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THE GEOLOGY OF THE BIRCH PORTAGE
BERYL PEGMATITE DEPOSIT, SASKATCHEWAN

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF MASTER OF SCIENCE

DEPARTMENT OF GEOLOGY

by

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MAY, 1964

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies for acceptance, a thesis entitled, "The Geology of the Birch Portage Beryl Pegmatite Deposit, Saskatchewan", submitted by Dennis Radcliffe, in partial fulfilment of the requirements for the degree of Master of Science.

ABSTRACT

The Birch Portage area of Saskatchewan consists of a series of low to medium grade Precambrian Kiseynew gneisses which have been intruded by acid and basic igneous rocks. These igneous bodies represent a series of magmatic injections into what are interpreted to be shear and extension fractures.

The field relationships, structural implications, petrography, mineralogy, and chemistry of the country and dyke rocks, have been examined for this hitherto uninvestigated area, and an original large scale field map is presented.

Studies on the thermal state of coexisting feldspars in the pegmatites and aplites indicates a depth of emplacement of approximately 9 miles. An X-ray diffraction pattern of beryl has been completely indexed, and these agree with the space group $C6/mcc$. The calculated specific gravities of beryls were found to be lower than the measured specific gravities. A fairly constant value of c_0 at $9.187 \pm 0.002 \text{ \AA}$ was determined for beryls, and it is thought that the variations are due to substitutions of larger atoms into the beryllium sites in tetrahedral coordination. The value of a_0 was found to vary from 9.201 to $9.255 \text{ \AA}$, and this expansion is due mainly to the introduction of " R_2O " elements, in excess of 2.5 mol. per cent into the open channels which parallel the 'c' crystallographic axis of beryl,

ACKNOWLEDGEMENTS

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FOREWORD

In 1962, the Saskatchewan Government, Department of Mineral Resources, Geological Survey Party #1, led by M.W. Pyke, mapped approximately 175 square miles of Saskatchewan Precambrian territory on a scale of 1" to 1 mile. This area was the Birch Portage, East Half, sheet number 63L/15 of the National Topographic Series. During the course of systematic mapping a series of beryliferous pegmatite dykes were discovered.

In 1963, Pyke mapped the Birch Portage, West Half, while the writer was employed as senior assistant. It was suggested by Pyke and chief geologist R. L. Cheesman, that detailed mapping of the beryl pegmatites might be worthwhile and the writer was thus able to spend 3 weeks mapping 4 square miles on a scale of 1" to 1/16 mile. Using this scale, it was possible to record every dyke, thicker than 2" (beryliferous and otherwise) on an outcrop map (see map 1 at end of thesis).

The distribution of beryl in the dykes while not necessarily haphazard, is definitely sporadic. Thus this current research has attempted to develop some means, structural, mineralogical, chemical or otherwise, of distinguishing barren dykes from beryliferous dykes, when the mineral beryl cannot be observed.

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CHAPTER I

INTRODUCTION

Location and Accessibility

The Birch Portage area of Saskatchewan is at latitude $54^{\circ}53'N$, longitude $102^{\circ}40'W$. It is situated 35 miles west-northwest of Flin Flon and 20 miles north of the Precambrian-Palaeozoic boundary (see Figure 1).

The area is conveniently demarcated by the north-south Sturgeon Weir River and pin pointed by the Birch Portage Indian Reservation which lies to the east of the river. Beryl pegmatites outcrop directly across from the reservation on the west bank of the river.

An all weather road links Flin Flon, Manitoba, with Prince Albert, Saskatchewan, and a northerly auxilliary road to Jan Lake passes two miles south of the deposit. The two miles can be covered by canoe from Leaf Rapids on Malign Lake which forms part of the Sturgeon Weir River System.

Dense bush covers most of this area but a burn 20 years ago cleaned the outcrops so that structures are easily visible. Spruce predominates with some poplar, birch, and stunted pine on the outcrops. A series of trapping trails traverse the area and link many of the outcrops.

Mapping Basis

The base map (Map 1) was derived from an R.C.A.F. vertical air photograph, number A 9788-43 (1946). This was enlarged to an approximate scale of $1/16$ mile or $330' = 1$ inch, Ground measurements by the writer and comparison with the National Topographic 1 mile map (63L/15) showed that the scale varied from 1 inch = $320'$ to $342'$. At the centre of the map the scale is 1 inch = $330'$. This decreases towards the

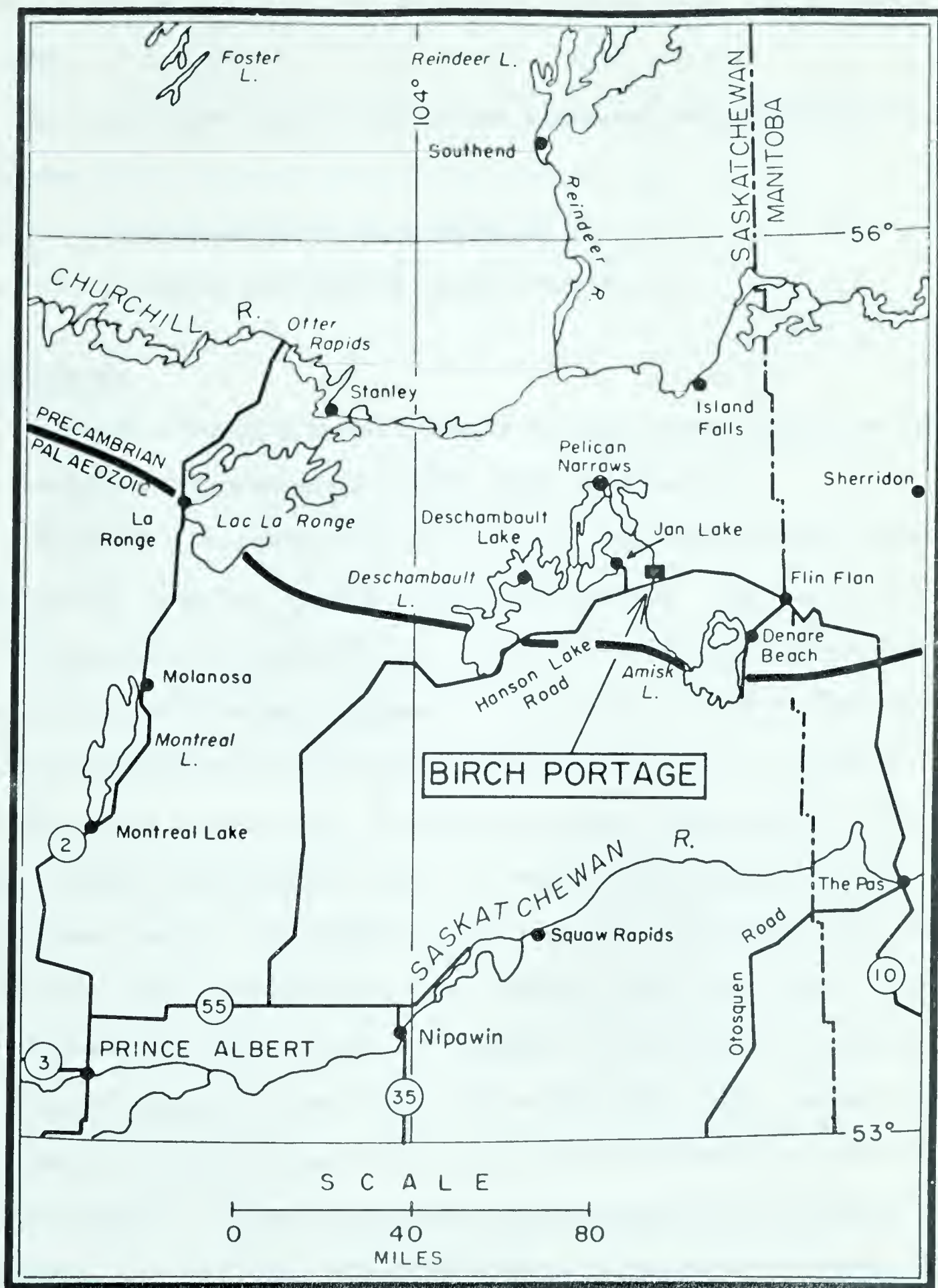


Figure 1. Location map for Birch Portage

corners to a maximum of $1'' = 342'$ and increases towards the centre sides to a minimum of $1'' = 320'$.

The area covered by the map is 4.3 square miles and the geographical co-ordinates are:

Latitude $54^{\circ}52'15''$ N. to $54^{\circ}54'15''$ N.

Longitude $102^{\circ}37'40''$ W. to $102^{\circ}40'20''$ W.

General Geology

The rocks of the south eastern portion of the Saskatchewan Precambrian Shield can be divided into three main groups: Amisk, Missi and Kiseynew.

The Amisk is the oldest group and consists of a sequence of volcanic rocks including dacites, andesites, rhyolites, tuffs and agglomerates. These are overlain by the Missi group consisting of sedimentary rocks, which are typically arkose, greywacke, argillite, quartzite and conglomerate. The youngest group is the Kiseynew, which consists of hornblende and/or biotite gneisses often extensively intruded by intermediate to acid igneous rocks. The Kiseynew gneisses which approach garnet amphibolite grade of metamorphism, may be the metamorphosed equivalent of the Missi and Amisk Groups. Byers (1954) describes the contact of Kiseynew rocks with Amisk and Missi rocks as conformable in some localities and faulted in other localities.

The area lies within the Churchill Geological Province of the Precambrian Shield and yield radiometric dates of from 1700 to 1900 million years. Lowdon (1961) reports an age of 1730 million years for the area northwest of Flin Flon and believes this to be a minimum. This age determination was done using the K-Ar method on primary biotites separated from a quartz diorite which is intrusive into the Amisk Group. The diorite is post-tectonic and is probably younger than the Missi group (Byers, 1954).

The map area of 4.3 square miles consists almost wholly of Kiseynew gneisses. These are hard, compact, foliated but not cleaved, hornblende and biotite gneisses,

generally with feldspar porphyroblasts. The rocks exhibit a general north-northeast strike and easterly dip of 50° . The eastern boundary of the map sheet is defined by the north-south trending Sturgeon Weir River marking the path of a probable fault system (Pyke, 1962). Movement along this fault may have been associated with the opening of a series of fractures of north-westerly strike and south-westerly dip. These fractures are filled by a series of dyke-like bodies of basic and acid igneous rocks consisting of a few metamorphosed diabases (?) and felsites, and many unmetamorphosed aplites and pegmatites.

Mapping by Pyke (1962) shows that these country rocks extend to the north and south of the present map sheet area and also occur to the east and northeast of the fault. To the southeast of the fault, Amisk- and Missi-type rocks outcrop. To the west the rocks assume a vertical or westerly dip and merge with a large granodiorite complex. Mapping by the writer shows that the beryl-bearing pegmatites are concentrated in the eastern half of the map immediately adjacent to the Sturgeon Weir fault.

The existence of wrench faults may be of some importance in localizing specific types of mineral deposits. To the east of the Sturgeon Weir Fault is the metalliferous rich region surrounding the mining town of Flin Flon. No beryl has been identified east of the fault. Fifteen miles west of the Sturgeon Weir Fault is another wrench fault (Tabenor Fault System) and it is only in between these two faults that beryl has been located. To date six occurrences of beryl have been identified and these are as follows:

- (1) Birch Portage Beryl Pegmatite Deposit is the largest and is described in this thesis. First discovered by Pyke (1962).
- (2) Seven Miles north of Birch Portage on the right bank of the Sturgeon Weir River (Pyke, 1962).

Longitude $102^{\circ}36'05''$ W. Latitude $54^{\circ}58'10''$ N.

- (3) One half mile west of the south-westerly tip of Onikup Lake (map 1). (Radcliffe, 1963).
- (4) Twelve miles southwest of Birch Portage, 1 mile north of Jackpine Lake (19 dykes) on the northwest bank of Hanson Lake (Pyke, 1963).
Longitude $102^{\circ}52'$ W. Latitude $54^{\circ}45'$ N.
- (5) Twenty-two miles northwest of Birch Portage, eight miles due east of Pelican Narrows, south of Wunehikun Bay (McKenzie-Noranda Prospector, 1963).
Longitude $102^{\circ}43'$ W. Latitude $55^{\circ}12'$ N.
- (6) Thirty-six miles north of Birch Portage, twelve miles north-northwest of Pelican Narrows, vicinity Ukoop Lake (Bear-Native of Pelican Narrows Indian Reservation, 1963).
Longitude $103^{\circ}05'$ W. Latitude $55^{\circ}20'$ N.

CHAPTER 2

STRUCTURE

Structure of the Country Rocks

The method employed for this study is based on that of described by Byers and Dahlstrom (1954). Their method was used as they have demonstrated that it works quite well for small areas in the Amisk-Wildnest area of Saskatchewan; Wildnest Lake lying some 10 miles east of Birch Portage.

The results of the study are shown on Fig. 2.a., where the structural elements have been projected onto the lower hemisphere of a Schmidt equal area stereographic net.

Byers and Dahlstrom define the axial line (hinge axis) as the intersection between the axial plane (axial surface) of a fold and the bedding plane (or foliation plane in recrystallized rocks). In the field this may manifest itself as a series of mineral lineations, within the plane of foliation. They define three mutually perpendicular tectonic axes; the 'b' axis being the axial line, the 'a' axis is the direction of tectonic transport, generally perpendicular to the axial line in cylindrical folds, and is manifested by small scale (parasitic) folding in the field; and the 'c' axis is perpendicular to the plane containing 'a' and 'b'.

On a stereonet the axial line appears as a point. Lineations in the 'a' direction are perpendicular to the 'b' axis (axial plane) and lie on a great circle perpendicular to the axial line. Thus since all lineations are measured in the plane of foliation then the 'a' lineation maximum must lie on the foliation pole circle, 90° distance from the corresponding limb maximum.

Accepting the interpretation of Byers and Dahlstrom (1954) as correct, then the plot (Fig. 2.a.) of the Birch Portage structural elements closely fulfills their

generalizations for cylindrical folds. The strike of the foliations varies between 355° and 45° and the dip between 30° to 70° E. The average is a strike of 16° and an easterly dip of 51° . Since these values fall within the highest density contour (Fig. 2.a.) for poles to foliation, they were accepted as the most frequently occurring combination of strike and dip. Contouring of these poles has shown a unimodal density distribution clearly demonstrating that only one limb of the fold is present in the area. That the other limb exists to the west with a westerly dip is known by personal observation and mapping by Pyke (1962). The term antiform rather than anticline is preferred for this structure as it is not known whether or not the structure is overturned.

The axial line plunges 30° toward the azimuth direction of 44° , indicating that the antiform has a north-easterly plunge. A vertical axial plane is indicated by the density distribution, but since only a statistically small number of stations were available (12) then the axial plane could have a dip within approximately 10° of vertical. The 'a' lineation maximum has a 36° plunge toward the azimuth direction of 159° .

Statistical analysis supports the field observations of an antiform with a north-easterly plunge. The compressive stress was thus probably in a N.W.-S.E. direction and under these conditions extension fractures could have opened up with long axis that paralleled the compressive stress axis. Fluid could have occupied these fractures and crystallized as the observed dyke-like bodies west of the Sturgeon Weir Fault.

Structure of the Aplites and Pegmatites

Poles to the attitudes of beryl pegmatites, barren pegmatites and aplites have been plotted on separate stereograms by projection onto the lower hemisphere of a Schmidt equal area net (Fig. 2, b.c.d.).

The density contours of the beryl pegmatites (Fig. 2.b.) indicates a general unimodal distribution whereas that of the barren pegmatites and aplites tend to indicate

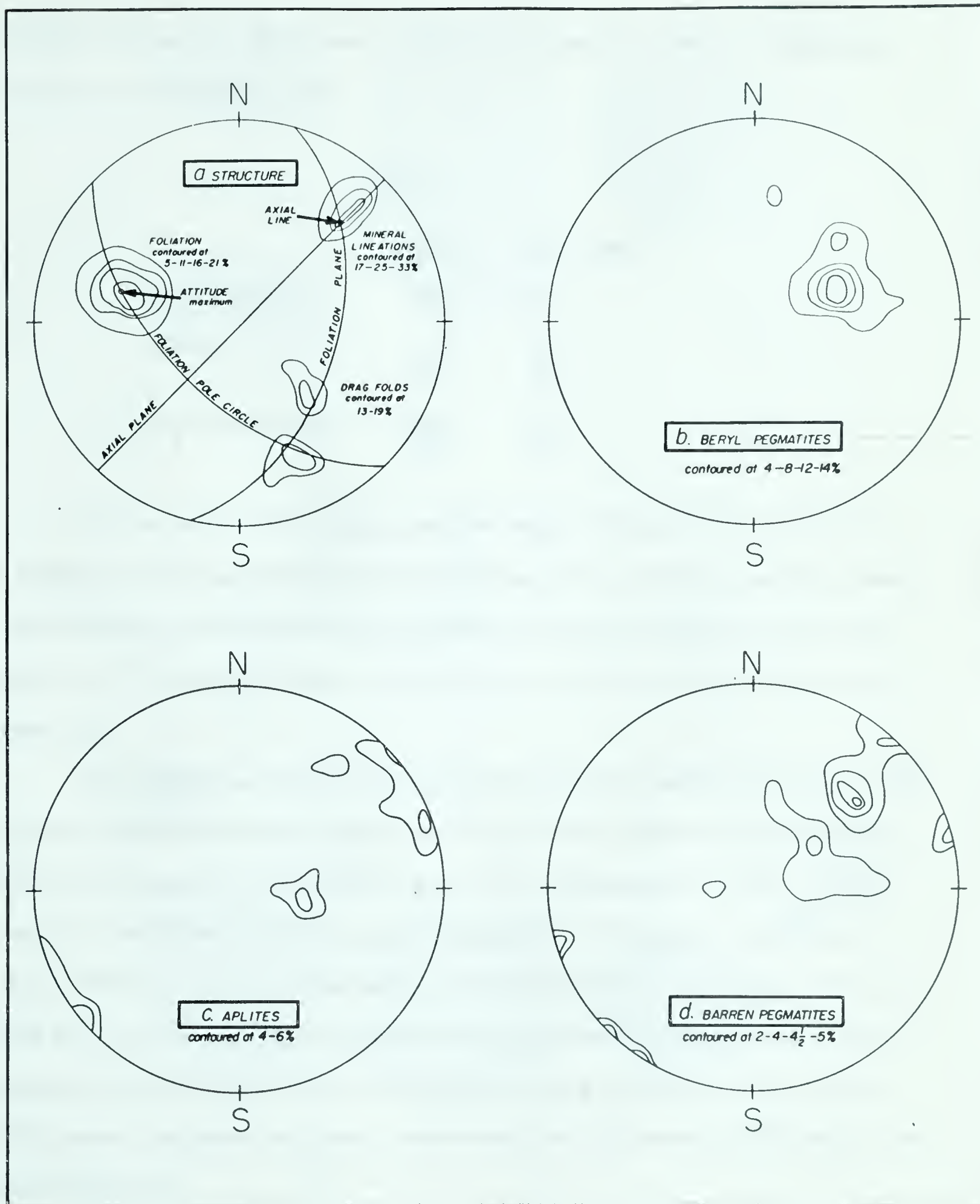


Figure 2. Contoured Structural Elements
of Birch Portage Rocks

a bimodal distribution. The contoured highs were accepted as the most frequently occurring attitudes and these are:

Table 1

	<u>Strike</u>	<u>Dip (west)</u>
Beryl pegmatites	335°	37°
Aplites	0°	28°
	329°	85°
Barren pegmatites	342°	26°
	315°	61°

On the basis of these figures, and the density distributions, it is difficult to distinguish beryl pegmatites from barren although the contouring interval indicates a closer grouping of attitudes for beryl pegmatites and a wide distribution for barren pegmatites. This may be a result, in part, of the restricted area of outcrop of beryl pegmatites.

The distribution of strike and dip about the most frequently occurring attitudes have been studied separately to see if there is any strike or dip particularly favoured by the beryl pegmatites. The standard deviation relationship of $1 \text{ S.D.} = \sqrt{\frac{\sum X^2}{N}}$ where X is the difference from the mean, computed for N samples. According to Spiegel (1961) $\pm 1 \text{ S.D.}$ from the mean includes 68.27% of all the points. It is unlikely that a normal distribution obtains and thus positive and negative values were computed separately using the most frequently occurring attitudes as "mean" values. In this manner the author was able to demonstrate that in all cases the distribution was skewed (Figure 3).

This investigation shows that, accepting a 2:1 probability ($\pm 1 \text{ S.D.}$), at best, the barren and beryl pegmatite dykes cannot be distinguished. However, Figure 3

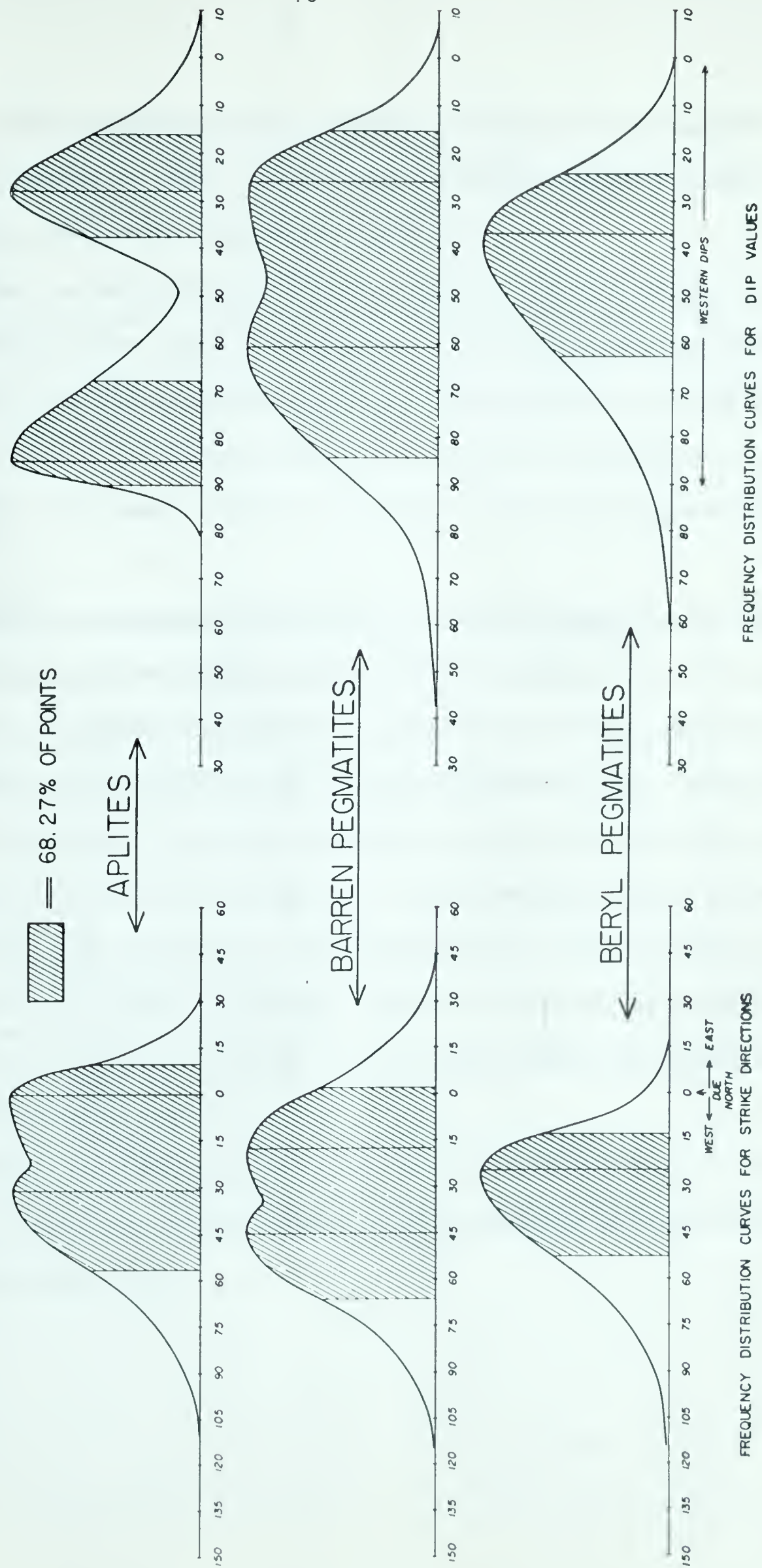


Figure 3. Statistical Analysis of the Strikes and Dips of Aplites and Pegmatites of the Birch Portage Beryl Pegmatite Deposit

demonstrates clearly the comparatively restricted distribution of beryl pegmatites. The distribution curves are useful in determining the possibilities of not locating beryl in a dyke outside of the range of 2:1 probability.

Perhaps the most significant fact to be gained from a study of these curves is the relationship between aplite and pegmatite. This distribution curves for the strike (Figure 3) show that aplites and pegmatites were intruded into a similar fracture system. The curves for the dips show clearly that aplites are concentrated in 2 sets of fractures, one at a high angle and one at a low angle with the beryl pegmatites lying in between.

A fan-like arrangement of fractures can thus be envisaged with the beryl pegmatite occupying the intermediate dips and aplites the high and low. There can be many theories to explain this phenomenon and an attractive theory would be that the intermediate dips were first formed and filled with granitic fluids. This would be essentially a closed system dynamically so that the solutions could cool slowly to form the coarse crystals characteristic of pegmatites. Later compressive forces further opened the pegmatite fractures and opened up additional fractures at higher and lower angles and these were actual leaks in the pressure chamber. Fluids would flow into these newer fractures and seal up the leakages. This would be indicated by the presence of bifurcating dykes (see Chapter 3).

A simple theory demands simultaneous entry of fluids into a fan-like fracture system when volatiles were able to escape more easily from high and low angle fractures compared to intermediate dips.

CHAPTER 3

PETROLOGY

Detailed mapping revealed the existence of eleven rock units in the area, which can be distinguished in hand specimen. These are arranged in order of increasing age based on field relationships:

Intrusive Rocks

- (11) Beryl Granite Pegmatite dykes
- (10) Barren Granite Pegmatite dykes
- (9) Granite Aplite dykes
- (8) Feldspar Porphyry dykes
- (7) Porphyritic Felsite dykes
- (6) Basic dykes

Country Rocks

- (5) Plagioclase Amphibole gneiss
- (4) Hornblende K-feldspar gneiss
- (3) Porphyroblastic Hornblende Plagioclase gneiss
- (2) Porphyroblastic Hornblende Biotite gneiss
- (1) Porphyroblastic Biotite gneiss

In thin section the feldspars of most rocks were seen to be generally untwinned and in some cases altered. The composition of the plagioclase feldspar was in most cases derived normatively with a knowledge of the mineralogy of the rock. While this method may not be fully satisfactory for the polymineralic country rocks it is thought to be quite adequate for the essentially trimineralic (quartz, 2 feldspars)

granitic rocks, especially since X-ray investigations on central phase feldspars shows only a small degree of solid solution.

The distinction of potassic and calcic feldspars necessary for modal analysis was made in some cases on the basis of colour staining, the potassic feldspars taking a yellow stain with sodium cobaltinitrite and the calcic feldspars a pink stain with erthyrosin "B" organic dye. By this method pure sodic feldspars remain unstained.

A. Country Rocks

The country rocks consist of hornblende and/or biotite gneisses arranged in a series of north-south subparallel bands with apparently conformable boundaries. The foliation in these rocks, which generally parallels the contacts, consists for the most part of alignment of the constituent minerals. Banding is rare and is restricted to the rocks bordering the Sturgeon Weir River. In the west the gneisses may locally lose their foliation and become granulitic, especially when the proportion of mafic minerals increases.

Feldspar porphyroblasts are a common feature of all the country rocks. In the east the tendency is for large (up to 3") discrete porphyroblasts whereas in the west the porphyroblasts are often joined up into stringers which parallel the foliation. In some localities advanced feldspathization together with the development of pygmatic folding justifies the term migmatite.

Three features are common to all of the country rocks and these are as follows:

- (i) All contain amphibolite boudins
- (ii) All contain amphibolitic layers
- (iii) All contacts are gradational

The gradational contacts of the rock units may occur along strike as well as across and this may indicate a facies change in the original sedimentary rocks (?) prior to metamorphism.

The north-south striking rocks have a general easterly dip near 50° , thus assuming no overturning the youngest country rocks outcrop in the east and the oldest in west. However in the northeast corner of the map sheet this generalization does not apply as rocks of all units are intimately associated with each other. Rocks in the eastern half of the map occasionally exhibit linaments of two types. Firstly small scale drag (parasitic) folds with a south-easterly plunge and secondly, north-easterly plunging mineral lineations. The interpretation of the structure of the area, discussed more fully in Chapter 2, is that of north-easterly plunging antiform whose eastern limb, which covers the map sheet, is truncated by a north-south wrench fault.

Porphyroblastic Biotite Gneiss (Unit I)

The porphyroblastic biotite gneisses are restricted to the northwest quadrant of the map where they embrace the Onikup Lake region. North of Onikup Lake the sequence is interrupted by an appearance of porphyroblastic hornblende biotite gneiss.

The porphyroblastic biotite gneisses are dark coloured, fine to medium grained rocks characterized by the presence of porphyroblastic alkali feldspars either as discrete crystals up to $1\frac{1}{2}$ " long when the rock is subgneissic, or as feldspar stringers ($1" \times 24"$) when the rock is highly gneissic. Fine aligned biotite flakes are the dominant mafic mineral and hornblende is apparently absent macroscopically; these are set in a matrix of fine dark coloured plagioclase feldspar. Banding is rare except in a locally developed migmatitic biotite gneiss, immediately south of Onikup Lake. Granulitic variants have also been recognized and their development is sporadic.

Thin section examination of a representative sample gave the following modal count: Quartz 12.3%, K-feldspar 3.4%, Na-feldspar 9.3%, Na-Ca feldspar 38.2%, biotite 23.0%, amphibole 12.6%, magnetite 1.11%. A few crystals of sphene also occur.

Texturally the rock is typically gneissic consisting of sub-aligned ferromagnesian minerals, often as discrete grains, in a matrix of anhedral, rather granular, interlocking

quartz and feldspar. The porphyroblasts consist of small potassic and sodic feldspars which are often twinned, the albite occasionally on combined albite and pericline laws giving the appearance of cross-hatching. Twinning is for the most part, secondary.

The soda-lime feldspars forming the bulk of the ground mass are completely untwinned and are essentially labradorites (An_{53} -norm.). Quartz is typically strained and often shows suture structure. Biotite is the most abundant mafic mineral and occurs mainly as disseminated flakes with irregular outlines. They are sometimes concentrated around the feldspar porphyroblasts. The amphibole content is interesting and was found to be tremolite-actinolite, its presence being confirmed by an X-ray diffraction pattern. It is characterized by very pale absorption colours from $\alpha < \gamma$ = colourless to γ = pale green and by partly corroded outlines.

The chemical composition of this rock is given in Table 2 and it may be noted that of all the country rocks analysed, the MgO content for the rock described above is the highest. This may be correlated with the appearance of tremolite-actinolite and perhaps the apparent absence of those types of amphiboles in the other country rocks i.e. the latter are characterized by hornblendes which have a lower MgO content than tremolite.

Porphyroblastic Hornblende Biotite Gneiss (Unit 2)

The porphyroblastic hornblende biotite gneisses occupy the southwestern quadrant of the map. They are essentially southerly a continuation of the porphyroblastic biotite gneisses and are distinguished by the presence of discrete hornblendes in a lighter coloured, more quartzitic matrix. The contact between the two units is along strike and gradational, perhaps indicative of a facies change in the original sedimentary rocks(?).

The macroscopic features and variations of the two rock types are essentially similar although the tendency for porphyroblastic stringers to develop appears to be higher in the south. East of Garnet Lake narrow (1/4") but very lengthy discontinuous

porphyroblastic stringers of feldspars, carrying quartz, characterize the large outcrop knob. These stringers parallel foliations and show that the plane of foliation is crenulated.

The modal analysis of a thin section gave: Quartz 26.2%, K-feldspar 4.7%, plagioclase 43.9%, biotite 10.3%, hornblende 11.4%. As before magnetite 0.49%, and sphene 3.03%, occur as accessory minerals.

Texturally the hornblende prisms occur as discrete, sub-aligned grains together with the biotite in an anhedral quartzo-feldspathic matrix. Quartz is dominant as interlocking stained grains exhibiting suture structures and the plagioclase (An_{21} - norm.) consists of untwinned anhedral oligoclase. The porphyroblasts in this rock consist wholly of potash feldspar which may or may not be twinned.

While the porphyroblastic hornblende biotite and biotite gneisses macroscopically appear to show a small along strike variation, mineralogically they are quite distinct. The porphyroblastic hornblende biotite gneisses have a much higher quartz content, the plagioclase is much more sodic and the amphibole is hornblendic rather than tremolitic. These changes are reflected chemically by a higher SiO_2 content, a lower MgO content and a higher Na_2O/CaO ratio.

Porphyroblastic Hornblende Plagioclase Gneiss (Unit 3)

The porphyroblastic hornblende plagioclase gneiss dominates the eastern half of the map sheet area where it occurs in two subparallel north-south bands. These bands 400 to 1700 feet wide are separated by a 400 foot band of hornblende K-feldspar gneiss (Unit 4) which is continuous down the map sheet area. Thus the porphyroblastic hornblende plagioclase gneiss can be divided into upper and lower units for the purpose of description of this present map sheet. The lower, most westerly unit, terminates in a southerly direction against porphyroblastic hornblende biotite gneisses but the nature of the contact was not observed. The upper unit is continuous down the map sheet area and in the northeast it outcrops on the west bank of the Sturgeon Weir River. The

porphyroblastic hornblende plagioclase gneisses can be identified topographically as they form prominent ridges, the rock itself being extremely tough and compact. It is noteworthy that beryl pegmatite dykes are concentrated in these rocks, and the significance of this fact is discussed later.

The porphyroblastic hornblende plagioclase gneisses are steel grey, hard, compact, coarse grained rocks, containing hornblende and feldspar porphyroblasts as discrete crystals. The sodium rich feldspar porphyroblasts average 2" in length, are not abundant in the rock. The discrete hornblende porphyroblasts average 3 x 2 mm in size. They are subaligned and define the foliation. The dark coloured hornblendes are set in an interlocking matrix of dark grey plagioclase feldspars.

Microscopic examination revealed the following mineralogy: Quartz 3.2%, K-feldspar 1.0%, plagioclase 77.1%, hornblende 7.3% and biotite 11.5%. No opaques or any other accessory mineral was identified in the slide under low and medium power microscope objectives.

Texturally the gneiss consists of anhedral crystals of all types, interlocked in a sub-aligned manner to produce a gneissic texture. There is some elongation of the untwinned plagioclase crystals of the groundmass and these consist of andesine (An_{39} -norm). The albite porphyroblasts generally displace the minerals of the groundmass and some sections show a small degree of twinning (Plate 1.A.). The hornblende porphyroblasts show a degree of alignment and their crystal form is quite irregular. They exhibit absorption colours which are typical for hornblende of medium grades of metamorphism i.e. α = yellow-green, β = green, γ = brown. Biotite is surprisingly abundant considering the macroscopic features of the rock. It occurs mainly as many very small flakes widely dispersed throughout the plagioclase groundmass. Quartz and K-feldspar are virtually absent.

Hornblende K-feldspar Gneiss (Unit 4)

The hornblende K-feldspar gneiss occurs as a band of rock little more than 400 to 600 feet wide, in the east part of the map, where it extends from the top of the map down to the bottom. It separates two units of the porphyroblastic hornblende plagioclase gneiss but does not usually form a prominent ridge. The contacts of the porphyroblastic hornblende K-feldspar gneiss are relatively sharp and the unit can normally be defined to within 24". It is noteworthy that this band is virtually devoid of dykes compared with the porphyroblastic hornblende plagioclase gneiss. The reason for this is not clear as both units are relatively tough and brittle and more susceptible to fracture than flow and are thus potential dyke hosts.

The hornblende K-feldspar gneiss is a coarse grained, compact rock, consisting of black discrete sub-aligned porphyroblastic hornblende grains set in a red-pink feldspar matrix and this colour relationship contrasts strongly with the dark coloured gneisses characterizing other rock units. Two variations occur, one being a fine grained red coloured gneiss consisting essentially of quartz and feldspar, and the other is amphibolitic and dark coloured. These two units form long narrow bands, little more than 10 feet wide but extending laterally for as much as 1000 feet.

The mode of the coarse grained porphyroblastic hornblende K-feldspar gneiss is as follows: Quartz 4.6%, K-feldspar 24.3%, albite 17.6%, plagioclase 28.6%, hornblende 14.4%. A few grains of magnetite and sphene occur as accessory minerals.

Porphyroblasts of hornblende and untwinned albite dominate the rock in thin section, the K-feldspar being restricted more in the groundmass. The soda-lime feldspar is probably andesine (An_{37} -norm). Quartz occurs as a few anhedral grains. Biotite exists as many very small dispersed flakes which give a rather high point count. Anhedral sphene (Plate 1.B.) is an accessory mineral.

Modal analysis of the fine grained red coloured quartz-feldspar gneiss showed the prevalence of quartz 26.6%, K-feldspar 29.6%, and plagioclase 41.4% and the virtual absence of mafic minerals, biotite constituting only 1.2% of the total volume.

Texturally a banded sequence can be recognized, consisting of 5 mm layers of potash feldspar and quartz, followed by potash feldspar and plagioclase. The latter is essentially oligoclase (An_{16} - norm). No porphyroblasts are in evidence and there is no obvious alignment of the essentially anhedral granular crystals (Plate 1.C.).

Plagioclase Amphibole Gneiss (Unit 5)

The plagioclase amphibole gneiss occupies the low lying ground of the west bank of the Sturgeon Weir River. It only extends halfway up the map sheet from the southern boundary.

It is characterized by bands of hornblende rich and hornblende poor layers, averaging 1" in width. The hornblende-rich layers appear to contain about 80% amphibole, justifying the term amphibolite; however the hornblende poor layers are quite rich in quartzo-feldspathic material.

The mode of the hornblende poor layers consists of quartz 31.3%, plagioclase 40.5%, hornblende 10.4%, biotite 16.9%, opaques 0.99%, and a little sphene occurs as an accessory mineral.

Texturally the small biotites occur in small sub-aligned flakes in a general decussate pattern, whereas the hornblende, which is porphyroblastic occurs as large irregular plates (Plate 2.A.) showing minimal orientation features. Anhedral quartz and plagioclases characterize the groundmass, some of the feldspar exhibiting albite twinning. An extinction angle determination on a section normal to (010) gave 20° indicating andesine, normative calculations indicate calcic andesine (Ab_{46}). The opaque minerals include yellow sulphide minerals.

Chemistry of the Country Rocks

A representative sample of each of the country rock units was analyzed during routine analysis of the granitic rocks. These are included in tabular form (Table 2) for the convenience of readers and future investigators.

With the exception of the hornblende K-feldspar gneiss (Unit 4), the chemical composition of the gneisses is very similar to the typical compositions of greywackes given by Pettijohn (1948, p. 250). It is interesting to note that Byers (1954) describes the frequent occurrence of greywackes to the east and southeast of Birch Portage.

Discussion of the Country Rocks

It is suggested that the Kiseynew gneisses, which outcrop throughout the map sheet area, are genetically paragneiss. The evidence for this conclusion is not overwhelming as no sedimentary structures have been preserved. The chemical composition alone is not conclusive evidence as greywackes being essentially unsorted sediments could reflect directly the chemical composition of an igneous rock from which they were mechanically derived. However the preponderance of metasediments to the east and southeast, the manner in which the foliation of the Birch Portage rocks follow contacts, the general nature of conformable boundaries, and the along strike variations and intercalations of horizons in the northeast, all suggest metamorphism of sedimentary rather than igneous rocks.

B. Intrusive Rocks

The intrusive rocks for the most part occupy fractures, and have a dyke-like form. They are generally less than 24" wide and less than 600 feet long. Contacts are normally sharp and straight except when the dyke has a shallow angle of dip. Termination of the dykes normally takes place over a distance of 30 feet and the overall picture of the dykes are as elongated lenticular envelopes.

TABLE 2
CHEMICAL ANALYSES OF GNEISSES

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>	<u>4a</u>	<u>5</u>
SiO ₂	61.45	64.00	64.90	57.52	73.16	63.92
TiO ₂	0.67	0.48	0.38	0.75	0.19	0.56
Al ₂ O ₃	13.25	13.02	13.57	13.57	13.54	13.08
Fe ₂ O ₃	6.79	6.82	5.86	7.44	1.53	7.97
MnO	0.10	0.09	0.08	0.10	0.03	0.09
MgO	4.60	2.98	2.80	3.10	1.86	3.40
CaO	5.62	4.30	4.58	4.85	1.19	5.45
Na ₂ O	3.17	3.70	3.65	5.55	3.68	3.26
K ₂ O	2.58	2.60	2.07	4.09	5.05	1.28
H ₂ O ⁺	?	?	?	?	?	?
H ₂ O ⁻	?	?	?	?	?	?
Others	0.08	0.07	0.06	0.15	0.13	0.07
Total	<u>98.23</u>	<u>98.06</u>	<u>97.95</u>	<u>97.32</u>	<u>100.36</u>	<u>99.08</u>

Trace Elements (ppm)

SrO	546	346	468	1237	1075	497
Li ₂ O	65	100	42	30	43	18
Rb ₂ O	43	46	43	87	74	15
Cs ₂ O	90	238	75	155	136	185

- 1 (560)---Porphyroblastic Biotite Gneiss
2 (587)---Porphyroblastic Hornblende Biotite Gneiss
3 (550)---Porphyroblastic Hornblende Plagioclase Gneiss
4 (527)---Coarse Hornblende K-feldspar Gneiss
4a (599)---Fine K-feldspar Gneiss
5 (598)---Plagioclase Amphibole Gneiss

The dykes are always discordant and there are very few mutually cross-cutting relationships considering that a total of 335 dykes were identified in an area not exceeding 4.3 square miles. Of the 335 dykes, 61 were beryl pegmatites, 186 were barren pegmatites (i.e. lacked visible beryl), 71 were granite aplites, 3 were feldspar porphyries, 5 were porphyritic felsites, and 9 were basic, perhaps metamorphosed diabbases.

The distribution of these various types of dykes shows barren pegmatites and granite aplites located throughout the map area. Feldspar porphyry dykes are located only in the extreme west, whereas felsites and basic dykes are restricted mainly in the east. Beryl pegmatites are concentrated to the east, southeast, southwest and west of Wells Lake and few beryl pegmatites were found more than 2,600 feet from the waters of the Sturgeon Weir River.

Structurally, the dykes have a northwest to west strike and have a dip from 0° to 90° southwest or south. Variations in the structural form can be observed, some of the granitic dykes bifurcating when the dips are intermediate. Occasionally intrusion has occurred into a steeply dipping conjugate fracture system, the area west of Wells Lake being a good example. When the dips are shallow the outcrop is irregular and dykes are often repeated by erosion e.g. west of Birch Rapids. Some of the shallowly dipping barren pegmatite dykes occupy a curved fracture pattern and may often have aplitic offshoots of steeper dips e.g. east of Garnet Lake. Significantly pegmatite dykes have been observed cross-cutting each other, indicating a series of injections e.g. east of Garnet Lake.

The age relationships of the various types of dykes was difficult to determine due to the general lack of cross-cutting relationships. However the basic dykes are cut by beryl pegmatite and show the highest degree of metamorphism and for these reasons are regarded as the oldest dyke rocks in the area. The felsites show a lower degree of metamorphism than the basic dykes and are often intimately associated with

them when fragments of the basic dykes are enclosed within the felsites. The felsites are thought to be post basic dykes in age. The feldspar porphyry dykes are unmetamorphosed and were placed next in the sequence especially since they are cross-cut by the younger granite aplites. Barren pegmatite crystallization probably followed an initial precipitation of aplite and as the crystallization of beryl was probably the last igneous event in the area, the beryl pegmatites have been given the youngest age.

Basic Dykes (Unit 6)

The basic dykes are generally restricted to the eastern portion of the map. They are generally less than 6 feet wide and usually 2 to 3 feet wide. Two distinct colours can be observed for the dykes, one being greenish black and the other, black. Both types resemble fine grained diabbases although the texture is rather sugary-granular. Yellow sulphides can occasionally be seen.

Structurally, the dykes have a fairly constant strike towards the northwest and are either vertical or dip 85° northeast. Often they are fracture cleaved parallel with the strike and in some localities e.g. east of Wells Lake, intrusion of felsite along a juxtapositional fracture has torn off and enclosed, fragments of the basic dykes.

Modal analysis of the greenish-black basic dykes gave the following results: orthoclase 17.4%, plagioclase 27.0%, tremolite 31.7%, and penninite 23.8%.

Texturally the rock consists of anhedral irregular tremolite crystals together with subhedral chlorite in a highly altered plagioclase matrix (Plate 2.B.). The texture of the black basic dyke is fairly similar but the mineralogy is a little different, consisting of biotite, magnetite, plagioclase and a few small hornblende porphyroblasts (Plate 2.C.). Both types of basic dykes tend to exhibit a decussate pattern of the mafic minerals when viewed under the microscope.

The mineralogy of the greenish-black dyke is unusual for metamorphosed basic dykes but all minerals were confirmed by an X-ray diffraction pattern. Chlorite gave a

fairly good $14 \text{ \AA}$ reflection and this fact together with the anomalous berlin blue interference colour suggests magnesium rich penninite. The amphibole is probably also magnesium rich (tremolite) as it shows very little pleochroism: $\alpha =$ colourless, $\beta =$ colourless?, $\gamma =$ pale green. Plagioclase feldspar is highly altered and its composition can only be estimated from the chemical composition, norm calculations indicating calcic andesine (An_{46}). The main alteration products of the feldspar is difficult to identify but is suspected to be kaolinite and not sericite as no $10 \text{ \AA}$ reflection appeared on the diffraction pattern. Kaolinite however is difficult to prove as its first order $7 \text{ \AA}$ (001) reflection is camouflaged by the chlorite second order (001) reflection. Theoretically one can distinguish dioctahedral kaolinite from trioctahedral chlorite by their respective (060) reflections, however all reflections in this region of the pattern were weak and indistinguishable.

The mineralogy is indeed unusual but is supported by chemical analysis (Table 3). The MgO content, for example, is 9.24% and this strongly supports the identification of magnesium rich tremolite and chlorite. The general chemistry except for the slightly high potash, and slightly low lime and alumina, is essentially that of a diabase. Thus the evidence of the general chemistry in addition to the occurrence of the metamorphic mineral tremolite, and the sugary, decussate appearance of the texture, strongly support the conclusion that these basic dykes are metamorphosed diabases, which may even have been potash metasomatised.

Porphyritic Felsite (Unit 7)

The porphyritic felsite dykes are restricted to the same general area as that of the basic dykes, although one notable exception outcrops east of the southern end of Onikup Lake. The attitudes of these dykes is the same as that of the basic dykes but they do not generally show the same high degree of fracture cleaving.

TABLE 3
CHEMICAL ANALYSES OF SOME DYKE ROCKS

	<u>6</u>	<u>7</u>	<u>8</u>
SiO ₂	52.80	71.02	72.24
TiO ₂	1.31	0.38	0.28
Al ₂ O ₃	11.00	13.90	13.90
Fe ₂ O ₃	10.99	2.97	2.25
MnO	0.14	0.04	0.03
MgO	9.24	---	---
CaO	7.13	1.21	1.29
Na ₂ O	2.37	3.28	3.31
K ₂ O	2.89	5.63	5.78
H ₂ O ⁺	?	?	?
H ₂ O ⁻	?	?	?
Others	0.14	0.07	0.07
Total	<u>98.01</u>	<u>98.50</u>	<u>99.15</u>

Trace Elements (ppm)

SrO	915	154	116
Li ₂ O	125	47	37
Rb ₂ O	140	382	370
Cs ₂ O	175	100	135

- 6 (530)---Metamorphosed Basic Dyke
 7 (515)---Porphyroblastic Felsite Dyke
 8 (503)---Feldspar Porphyry Dyke

ANALYST: Dennis Radcliffe

The felsite is a dark pink-grey colour and is prominently flow banded parallel to the dyke walls. The large euhedral K-feldspar (5 x 2 mm) crystals are similarly oriented parallel with the dyke walls and for this reason they are thought to be original phenocrysts rather than porphyroblasts.

Modal analysis revealed quartz 8.3%, orthoclase 51.0%, plagioclase 29.7%, biotite 7.2% and accessory magnetite 2.4%.

Quartz exists as discrete grains moulded by the ill-defined potash and soda rich feldspars of the groundmass. Norm calculations of the plagioclase indicates oligoclase (An_{18}). Biotite occurs mainly as small oriented flakes but a few larger flakes (2 mm) may show alteration to magnetite. The K-feldspar phenocrysts do not usually poikilitically enclose other minerals but a few anhedral crystals do so and this may indicate a small development of porphyroblasts. This feature together with the altering biotite may be indicative of a small degree of metamorphism.

Feldspar Porphyry Dykes (Unit 8)

Only 3 narrow dykes of feldspar porphyry have been observed and these are located on one outcrop knob on the western boundary of the map immediately south of Onikup Lake. One of the dykes is cut by granite aplite.

The feldspar porphyry dykes are a white-grey colour consisting of large (15 x 5 mm) white orthoclase phenocrysts in random orientation surrounded by small biotites in a matrix of plagioclase and quartz.

Modal analysis showed quartz 18.2%, orthoclase 39.6%, plagioclase 36.8%, and biotite 5.5%.

The orthoclase phenocrysts form euhedral crystals whereas the quartz of the groundmass is quite anhedral. The plagioclase feldspars, which are untwinned, were calculated from the chemical composition to be oligoclase (An_{19}). Biotite flakes appeared to have no preferred orientation.

The chemical analysis (Table 3) of this rock showed that it has essentially an acid alkali rich granite composition, but there is no apparent correlation with this rock type and the later granite aplites and pegmatites. Perhaps of interest is the essentially similar chemical composition of the porphyritic felsites and feldspar porphyries although they are quite dissimilar in colour, structure, location, grain size and metamorphism. Perhaps the difference in metamorphism is explained by the differences in the location.

Aplite Dykes (Unit 9)

The aplite dykes are usually found in association with the barren and beryl pegmatites and are distributed throughout the map area. They are generally 6" to 12" wide, and occur mainly as individual units although in a few bifurcating dykes one limb is occupied by pegmatite and the other by aplite. However in a few cases multiphase dykes occur consisting of 4" to 6" wide alternating bands of pegmatite and aplite. The aplites are fine grained and usually either cream or red consisting essentially of quartz and feldspar and may contain euhedral magnetite.

Structurally the aplites are interesting and frequency distribution diagrams for the dip of the dykes (Figure 3) shows a marked tendency for shallow or steep dips. The strike of the dykes is similar to that of the pegmatites.

Thin section analysis of red and cream aplites yielded a monotonous assemblage of quartz, K-feldspar, albite and biotite, with accessory magnetite. Four specimens were modally analyzed and these results are given below in Table 4.

TABLE 4

	<u>Colour</u>	<u>Quartz</u>	<u>K-Fsp</u>	<u>Albite</u>	<u>Biotite</u>	<u>Magnetite</u>
Cream	(501)	26.0	30.0	38.3	5.4	-
Cream	(519)	41.5	32.5	23.3	2.3	-
Red	(510)	22.5	27.8	46.8	1.8	1.1
Red	(533)	28.9	21.6	40.2	7.5	1.8
Multiphase	(537)	22.1	16.0	60.0	1.3	0.2

Texturally, the aplites are granular to sugary and a few feldspar phenocrysts can be observed. These latter consist mainly of untwinned K-feldspar. Albite may form smaller phenocrysts and these are generally highly twinned according to the combined albite and carlsbad or combined albite and pericline laws (Plate 3.A, B, and C). No microperthite was observed.

The matrix is difficult to analyze as quartz, which is anhedral, is closely associated with feldspar. Staining of the slides shows that the potash and soda feldspars are intimately associated with each other and there is only a partial separation to the two phases. Euhedral magnetite and flakes of biotite can be observed.

The red aplites are distinguished mineralogically from the cream aplites by the existence of magnetite and by a higher albite and lower orthoclase content, perhaps the inverse relationship implied by the colour of the dykes.

The mineralogy of the multiphase dykes is similar to the aplites but the higher albite and lower orthoclase content suggests the influence of the pegmatitic material as slow cooling granitic compositions become increasingly enriched in soda and relatively depleted in potash.

Chemical analysis of these rocks (Table 8) supports the observed modal analysis and these are given below, in part:

TABLE 5

	<u>K₂O</u>	<u>Na₂O</u>
Cream (501)	4.93	3.62
Cream (519)	6.42	2.87
Red (510)	4.22	4.06
Red (533)	3.89	4.44
Multiphase (537)	3.25	5.39

The general chemical composition of the aplites is essentially the same as the pegmatites suggesting that the aplites and pegmatites are of the same parent material which was apparently highly differentiated prior to intrusion (see Differentiation Index).

Barren Pegmatite (Unit 10)

The barren pegmatites may be dyke-like or stock-like bodies. The stocks are massive and discordant, generally outcropping as a series of small knobs averaging 100 feet in diameter. They consist of coarse red perthite crystals often 2" long, quartz, biotite and muscovite. They are restricted to the northwest and southeast of Onikup Lake and the absence of distorted foliations in the country rocks suggests that these pegmatites may be a replacement type.

The dyke-like bodies of barren pegmatite are distributed throughout the map area. They are only distinguished from beryl pegmatites by the absence of beryl. They do not usually exceed 3 feet in width and can be traced for as much as 2000 feet along strike. Quartz and 1" feldspar crystals characterize these dykes, the feldspar being red, pink, cream or white. Chilled contacts are generally less than 1/4" wide. Quartz, albite and K-feldspar can be seen to be co-precipitated from the margins of the dykes up to the central phase, indicative of cotectic, subsolvus crystallization.

In general two distinct structural assemblages can be observed (see Plate 7). These are firstly a rhythmic alternation of bands parallel with the dyke walls consisting essentially of fine albite and quartz followed by K-feldspar and quartz. Secondly coarse K-feldspar, albite and quartz may grow perpendicular to the dyke walls in discrete subparallel crystals; occasionally the central phases of such dykes may grow in a plane parallel with the dyke walls.

Coarse feldspar crystals measuring up to 2" long often occur in the central phases of the dykes and there are usually perthitic, X-ray investigations showing only 5% albite remaining in solid solution with the potassic phase. Evidence for replacement perthitization is small.

Four barren pegmatites have been examined microscopically and chemically. These specimens were sampled from 8" dykes, the 4" hand specimen being regarded as representative of the whole dyke (also for beryl pegmatites). The modal analysis of the four barren pegmatites is given below in Table 6.

TABLE 6

	<u>526</u>	<u>531</u>	<u>539</u>	<u>556</u>
Quartz	31.4	23.5	20.8	26.8
K-Fsp.	13.2	14.0	23.1	13.1
Plagioclase	53.6	62.0	46.1	58.8
Biotite	0.9	0.5	6.7	0.3
Muscovite	--	Trace	--	--
Garnet	0.8	--	0.8	--
Apatite	Trace	--	Trace	--
Magnetite	0.2	--	2.4	1.1
Plag. Content	Ab ₉₃	Ab ₉₄	Ab ₉₀	Ab ₉₆

The general mineralogy and chemistry of the barren pegmatite is essentially similar to that of the beryl pegmatites and a fuller discussion is given in that section.

Of particular interest in the barren pegmatites is the twinning in feldspars, some of central phase K-feldspars being twinned on the combined albite and pericline laws to give the characteristic "cross hatched" effect, characteristic of microcline. Plate 4A shows two microclines, one of which is partly replaced by quartz. The albites are usually highly twinned and on the basis of criteria advanced by Vance (1961) exhibit excellent examples of secondary glide twinning (Plate 4.C.). In general, however, twinning is not common in the feldspars and in some dykes (e.g. D.R. 531), the potash and soda rich phases are only partly unmixed.

Beryl Pegmatite Dykes (Unit 11)

In general the beryl pegmatites, which always occur in dyke-like bodies, are restricted to the eastern portion of the map where they roughly parallel aplites and barren pegmatites in a northwesterly trend and intermediate south-westerly dip. Approximately 90% of the beryl pegmatite dykes are restricted to a zone extending no further than 2,600 feet west of the Sturgeon Weir River. The area of greatest concentration is immediately to the east, west and south of Wells Lake.

Structurally, the dykes lie on the eastern limb of a northeasterly plunging antiform which has been truncated by transcurrent faulting, the dykes being concentrated near the fault zone (compare Beus, 1962, p. 46). The dykes appear to occupy original extension fractures.

The granite beryl pegmatites are with the exception of beryl, mineralogically of the simple type (Turner and Verhoogen, 1960). The dykes are generally no more than 24" wide with many only 3" to 6" wide. They have sharp, straight contacts and are continuous for little more than 600 feet along strike. A simple zonation can be sometimes recognized, many of the dykes consisting of quartz and dominantly feldspar

with a central zone of blocky quartz. Beryl is sometimes located at the junction of these two zones, while in other dykes in which blocky quartz is absent, beryl occupies the central zone.

Beryl occurs in small pockets averaging approximately 12" square by 1" thick, and the greatest dimension of these pods parallels the plane of the dykes. The highest concentration of beryl was 12, 1" crystals in 1 square foot which is about a 10% concentration. In such a dyke only 1 pocket of beryl might be visible out of a possible 600 feet outcrop length.

The Birch Portage beryls are always opaque, and green or yellow-green in colour. They are sometimes zoned from green to yellow toward the outside of the crystal, and this occurs particularly when the beryl is in close association with feldspar. The colour change thus appears to be influenced by the chemical constituents of the feldspars, in particular the alkalis. The largest beryl crystal observed was 1" in diameter and of undeterminable length. Commonly crystals are 1/2" to 1" long and 1/4" to 1/2" in width.

Beryl is found in association with quartz, K-feldspar, albite or any combination of the three minerals. It is also found in coarse grained dykes (1" feldspars) dominantly fine grained dykes (1/4" feldspars), wide dykes (36") or narrow dykes (6"), and in red, pink, and cream coloured dykes. Of the 61 beryl pegmatite dykes examined, all contained feldspars and quartz, and in addition to these minerals, 4 dykes contained magnetite, 6 dykes contained garnet, 6 dykes contained biotite and 8 dykes contained muscovite. Columbite-tantalite (1 dyke) and monazite (1 dyke) have also been reported (Pyke, 1963, personal communication). In fact, the beryl pegmatites show a greater mineralogical and chemical variation than the other granitic rocks in the area. It would appear therefore that the pegmatite containing beryl is only a host rock which had little to do with an in situ concentration and crystallization of beryl.

In hand specimen, structures identical to those described for barren pegmatites can be observed (see Plate 7) i.e. parallel growth of crystals from the dyke margins and rhythmic repetition of bands parallel to the plane of the dyke. The following bands were identified (Plate 7, A) by comparison with the thin section.

- (1) Marginal phase. Absent
- (2) Coarse, K-feldspar + Quartz + Albite
- (3) Fine, Albite + Quartz + little K-feldspar
- (4) Coarse, K-feldspar + Quartz + Albite
- (5) Fine, Albite + Quartz + little K-feldspar + Beryl

Parallel growth of quartz, K-feldspar and albite from the dyke walls toward the central phase is exemplified by Plate 7, B. The conclusion reached on this evidence is the same as that for the barren dykes that is, the cotectic, subsolvus crystallization of feldspars and quartz.

The modal analysis of four representative beryl pegmatites is given below in Table 7.

TABLE 7

	<u>512</u>	<u>521</u>	<u>522</u>	<u>601</u>
Quartz	38.3	18.7	30.1	24.2
K-feldspar	3.7	41.7	28.1	30.9
Albite	55.4	34.5	39.7	42.8
Biotite	2.0	--	0.9	--
Muscovite	--	--	Trace	Trace
Garnet	Trace	5.2	--	--
Apatite	Trace	--	--	--
Magnetite	0.7	--	1.2	--
Beryl	Trace	Trace	--	2.1
Plagioclase content	Ab ₉₁	Ab ₈₉	Ab ₉₄	Ab ₉₆

Quartz is always prevalent in the beryl pegmatites and is often a reddish colour. It may show suture texture and straining near the margins of the dykes (Plate 5.A.). Garnet is an accessory mineral and is most probably the spessartite variety (Plate 5.B.) as garnetiferous pegmatites are characterized by a high MnO content, often higher than that of the surrounding country rocks.

Beryl, when viewed under high power, could be seen to contain a large number of inclusions, many of which could be fluid inclusions with high birefringence (Plate 5.C) (These average $1/50$ mm.).

The feldspars are mainly untwinned, albite twinning of albite being the more prevalent especially in perthites. These microperthites (Plate 6) usually exhibit exsolved albite lenticles averaging 0.3×6 mm. Plate 6.C shows untwinned exsolved albite and Plate 6.B demonstrates a type of perthite in which the twinned exsolution lamellae have been irregularly extended by some replacement perthite. The small degree of replacement in the pegmatites dykes is unusual as many beryl pegmatites show advanced replacement features (Beus, 1962) including, microcline perthites, cleavelandites, and muscovites.

Most authors describe either albite or microcline as the feldspar commonly associated with beryl. The potash feldspars from the central phases of beryl pegmatites are untwinned and it was necessary to use X-ray means to determine whether or not the feldspar was triclinic (microcline) or monoclinic (orthoclase). Using the $\Delta^{\circ}2\theta_{131-1\bar{3}1}$ (MacKenzie, 1954) a medium to high triclinicity of $0.76 - 0.79$ was determined for 4 feldspar crystals. Maximum microcline has χ equal to $87^{\circ}30'$ giving a $\Delta^{\circ}2\theta$ of 1.0 . These feldspars are thus triclinic and correctly described as untwinned microcline.

Discussion of the Pegmatite Petrology

The orientation of the crystals and the crystal assemblages are of great genetic significance in the pegmatites. The manner of the parallel growth of crystals perpendicular from the dyke walls, implies a liquid central phase through which migrating solutions could pass. That this occurred is strongly suggested by the composition of the rhythmic banding as liquids in the ternary system $\text{SiO}_2\text{-NaAlSi}_3\text{O}_8\text{-KAlSi}_3\text{O}_8$ can never pass from one feldspar field to the other, either under equilibrium or disequilibrium crystallization, in a closed system.

The passage of solutions through the still liquid inner portions of the crystallizing dykes collected and concentrated the beryllium into discrete pockets where it crystallized. These migrations thus explain why there is no correlation between the beryl and the rest of the pegmatite dyke. Thus the pegmatite is only a host rock for the migratory beryllium rich solutions. The evidences of the fluid inclusions in beryl is explained by Beus (1962) and Cameron (1953) in terms of a beryllium-alkali-volatile complex, probably an immiscible phase, migrating into loci of pressure lows, the crystallization of beryl trapping some of the original fluid complex.

The evidence of co-precipitated feldspars can be interpreted on the basis of experimental work (Tuttle and Bowen, 1958) to imply crystallization at high pressures, and at a temperature little in excess of 660°C . At this temperature monoclinic potash feldspar (orthoclase) is precipitated; however the central phase potash feldspars are quite highly triclinic (microcline) and readjustment must have occurred at some lower temperature. Smith (1960) indicates that this structural inversion, involving the ordering of silicon and aluminum, may occur at a temperature less than 300°C .

Chemistry of the Aplites and Pegmatites

The chemical analyses of four representative samples of aplites, barren pegmatites and beryl pegmatites are given in Tables 8, 9, and 10, respectively.

TABLE 8
CHEMICAL ANALYSES OF APLITES

	<u>501</u>	<u>510</u>	<u>519</u>	<u>533</u>
SiO ₂	75.01	74.02	74.40	74.08
TiO ₂	0.05	0.08	0.04	0.03
Al ₂ O ₃	13.97	13.10	14.47	13.92
Fe ₂ O ₃	1.39	2.41	0.73	1.40
MnO	0.03	0.03	0.03	0.11
MgO	---	---	---	0.96
CaO	0.75	1.60	0.98	0.76
Na ₂ O	3.62	4.06	2.87	4.44
K ₂ O	4.93	4.22	6.42	3.89
H ₂ O ⁺	?	?	0.04	0.26
H ₂ O ⁻	?	?	0.53	0.28
Others	0.04	0.05	0.05	0.05
Total	<u>99.79</u>	<u>99.57</u>	<u>100.56</u>	<u>100.18</u>
<u>Trace Elements (ppm)</u>				
SrO	43	216	85	61
Li ₂ O	24	29	43	37
Rb ₂ O	210	113	204	277
Cs ₂ O	126	107	147	159

ANALYST: Dennis Radcliffe

TABLE 9
CHEMICAL ANALYSES OF BARREN PEGMATITES

	<u>531</u>	<u>539</u>	<u>540</u>	<u>556</u>
SiO ₂	72.98	76.28	77.40	75.32
TiO ₂	0.01	0.02	0.01	0.01
Al ₂ O ₃	15.17	13.61	13.73	14.08
Fe ₂ O ₃	0.48	1.25	0.68	0.83
MnO	0.06	0.39	0.12	0.04
MgO	---	---	---	---
CaO	0.65	0.48	0.26	0.51
Na ₂ O	5.39	4.68	3.82	4.28
K ₂ O	3.29	3.31	3.72	3.68
H ₂ O ⁺	---	?	---	?
H ₂ O ⁻	0.30	?	0.35	?
Others	0.04	0.04	0.06	0.05
Total	<u>98.37</u>	<u>100.06</u>	<u>100.15</u>	<u>98.80</u>

Trace Elements (ppm)

SrO	35	24	19	41
Li ₂ O	31	108	29	13
Rb ₂ O	188	191	359	325
Cs ₂ O	181	110	172	123

ANALYST: Dennis Radcliffe

TABLE 10
CHEMICAL ANALYSES OF BERYL PEGMATITES

	<u>512</u>	<u>521</u>	<u>522</u>	<u>601</u>
SiO ₂	71.08	71.34	74.36	73.26
TiO ₂	0.02	0.02	0.04	0.01
Al ₂ O ₃	15.65	15.09	14.35	14.51
Fe ₂ O ₃	1.28	1.80	1.23	0.45
MnO	0.19	0.78	0.09	0.07
MgO	---	---	---	---
CaO	1.04	0.62	0.63	0.54
Na ₂ O	7.72	3.60	4.02	3.63
K ₂ O	0.77	6.07	3.81	6.13
H ₂ O ⁺	0.22	0.47	?	?
H ₂ O ⁻	0.23	0.46	?	?
Others	0.02	0.08	0.05	0.08
Total	<u>98.22</u>	<u>100.33</u>	<u>98.38</u>	<u>98.68</u>

Trace Elements (ppm)

SrO	72	54	65	57
Li ₂ O	16	39	28	26
Rb ₂ O	55	469	315	515
Cs ₂ O	73	238	135	233

ANALYST: Dennis Radcliffe

TABLE 11
CHEMICAL ANALYSES OF ALKALI FELDSPARS

	<u>549</u>	<u>562</u>	<u>605</u>	<u>608</u>
SiO ₂	67.64	65.62	68.92	65.00
TiO ₂	---	---	0.01	0.03
Al ₂ O ₃	18.29	18.48	17.71	17.24
Fe ₂ O ₃	0.43	0.14	0.26	1.08
MnO	0.15	0.01	0.01	0.02
MgO	---	---	---	---
CaO	0.20	0.06	0.08	0.08
Na ₂ O	5.71	3.27	3.37	3.02
K ₂ O	6.30	11.95	10.17	12.00
H ₂ O ⁺	?	?	?	?
H ₂ O ⁻	?	?	?	?
Others	0.09	0.15	0.14	0.14
Total	<u>98.81</u>	<u>99.68</u>	<u>100.67</u>	<u>98.61</u>

Trace Elements (ppm)

SrO	28	33	22	25
Li ₂ O	34	28	38	45
Rb ₂ O	652	1166	972	989
Cs ₂ O	226	299	388	314
TRICLINICITY	0.78	0.76	0.76	0.76
Or CONTENT	95	96	95	93
Ab CONTENT	98	99	99	99

549. Pink Or-Ab intergrowth (with spessartite) from Beryl Pegmatite

562. Blue-Grey Perthite from Beryl Pegmatite

605. Red Perthite from Beryl Pegmatite

608. Red Perthite from Barren Pegmatite

ANALYST: Dennis Radcliffe

Four perthitic feldspars, sampled from the central phase of the pegmatite dykes, have also been analyzed (Table 11).

The MnO content of the pegmatites and aplites is on the average higher than that of the country rocks. A direct correlation can be made with the MnO content and volume of garnet in the modal analysis. For this reason the garnets are thought to consist predominantly of spessartite. The garnet in some cases appear to be a primary mineral and the reason for the rather high MnO content in these highly differentiated rocks is not known.

It is exceedingly difficult to recognize any significant chemical differences which would distinguish beryl from barren pegmatites. Small differences exist, for example, the silica content is a little higher and alumina content a little lower in the beryl pegmatites. The strontium and cesium contents and the Sr/Ca ratio appear to be a little higher also. Comparison with the aplites shows however that these differences are not real.

Perhaps of significance is the very similar chemistry of aplites and pegmatites strongly indicating that the melt from which these rocks crystallized was highly differentiated prior to intrusion. A measure of the degree of differentiation is afforded by the Differentiation Index which was introduced by Thornton and Tuttle (1960).

6.3 Differentiation Index

The Differentiation Index (D.I.) is derived by calculating an anhydrous norm and summing the minerals which fall into "Petrogeny's Residua System" i.e. the system silica-nepheline-kalsilite. Since the Birch Portage acid dykes are saturated rocks the Index was obtained by summing normative quartz, orthoclase and albite.

The theory of the index is that all complex magmas produce liquids which move toward, and for all practical purposes, eventually reach the system SiO_2 - NaAlSiO_4 - KAlSiO_4 . This has been corroborated by recent experimental studies (e.g. Tuttle and Bowen, 1958).

A rock consists wholly of anorthite (ideal anorthosite) would have a D.I. of zero, whereas a rock containing pure albite, or orthoclase, or quartz, or all three, would have a D.I. of 100. It is possible to distinguish saturated, undersaturated and oversaturated rocks. Since orthoclase contains the highest silica content, then any composition above the Or-An join on the plot of D.I. against SiO_2 would be oversaturated. A composition falling below the Ab-An join would be undersaturated and any falling between the Ab-An and the Or-An joins would be saturated.

On the basis of these principles the Birch Portage pegmatites and aplites are oversaturated rocks with a Differentiation Index generally in excess of 90. The calculated Differentiation Indices for the chemically analyzed samples given in Tables 8, 9, and 10, are given below in order of appearance in their respective Tables.

TABLE 12

Aplites	86	90	90	93
Barren Pegmatites	90	94	96	96
Beryl Pegmatites	89	91	94	96

The aplites are only slightly more basic than the pegmatites indicating that some fractionation occurred during pegmatite crystallization. The beryl pegmatites show a wider variation than the barren pegmatites, and if anything, are more basic. This provides more evidence that the beryl and barren pegmatites are for all practical purposes the same rock and that the presence of beryl in the pegmatite does not necessarily indicate a different crystallization history for the bulk of the minerals in the pegmatites.

It might be added that the plots of D.I. against all the experimentally determined oxides agree well with the graphs presented by Thornton and Tuttle (1960) who plotted 5,000 analyses from Washington's Tables.

Petrogenesis of the Aplites and Pegmatites

The thermal and dynamic history of these rocks can be further understood by considering the aplites and pegmatites in terms of "Petrogeny's Residua System". This is defined as the system, $\text{SiO}_2\text{-KAlSiO}_4\text{-NaAlSiO}_4$.

In this study only the field of saturated rocks need be considered and discussion is confined to that part of the system defined by the three phase triangle, $\text{SiO}_2\text{-KAlSi}_3\text{O}_8\text{-NaAlSi}_3\text{O}_8$. This is frequently referred to as the granite system.

Normative quartz, orthoclase, and albite were recalculated to 100% for the aplites, beryl pegmatites, barren pegmatites and two multiphase dykes (Figure 4.A.). As might be expected there was no area restricted to any particular type of rock and it has already been pointed out that the liquid from which these rocks crystallized was probably highly differentiated prior to intrusion. The points were contoured (Figure 4.B), so that the density distribution of points could be examined. Two trends now become obvious. Firstly, a high exists at $\text{Q}_{36}\text{Ab}_{39}\text{Or}_{25}$ and there is an oval density distribution about this high indicating a distribution of compositions with fixed Or_{25} contents and varying Ab-Q. This trend is the same as that described by Tuttle (1958, p. 79) for 571 analyzed plutonic rocks in Washington's Tables that carried a differentiation index greater than 80. One important difference exists in that in Tuttle's data the potash feldspar content is Or_{30} which is 5% higher than for Birch Portage rocks. This demonstrates that Birch Portage granitic rocks are 5% richer in albite (and quartz) relative to orthoclase and provides additional support for the author's contention that these rocks are highly differentiated. Since sodic feldspar has the lowest melting point of all common rock forming minerals, then liquids in a fractionating sequence would become increasingly enriched in albite.

Secondly there is another density distribution from the contoured high toward the potash rich feldspar end member. This is difficult to explain. Bowen (1937) clearly demonstrated that a "sink" exists in the system into which all fractioning liquids should

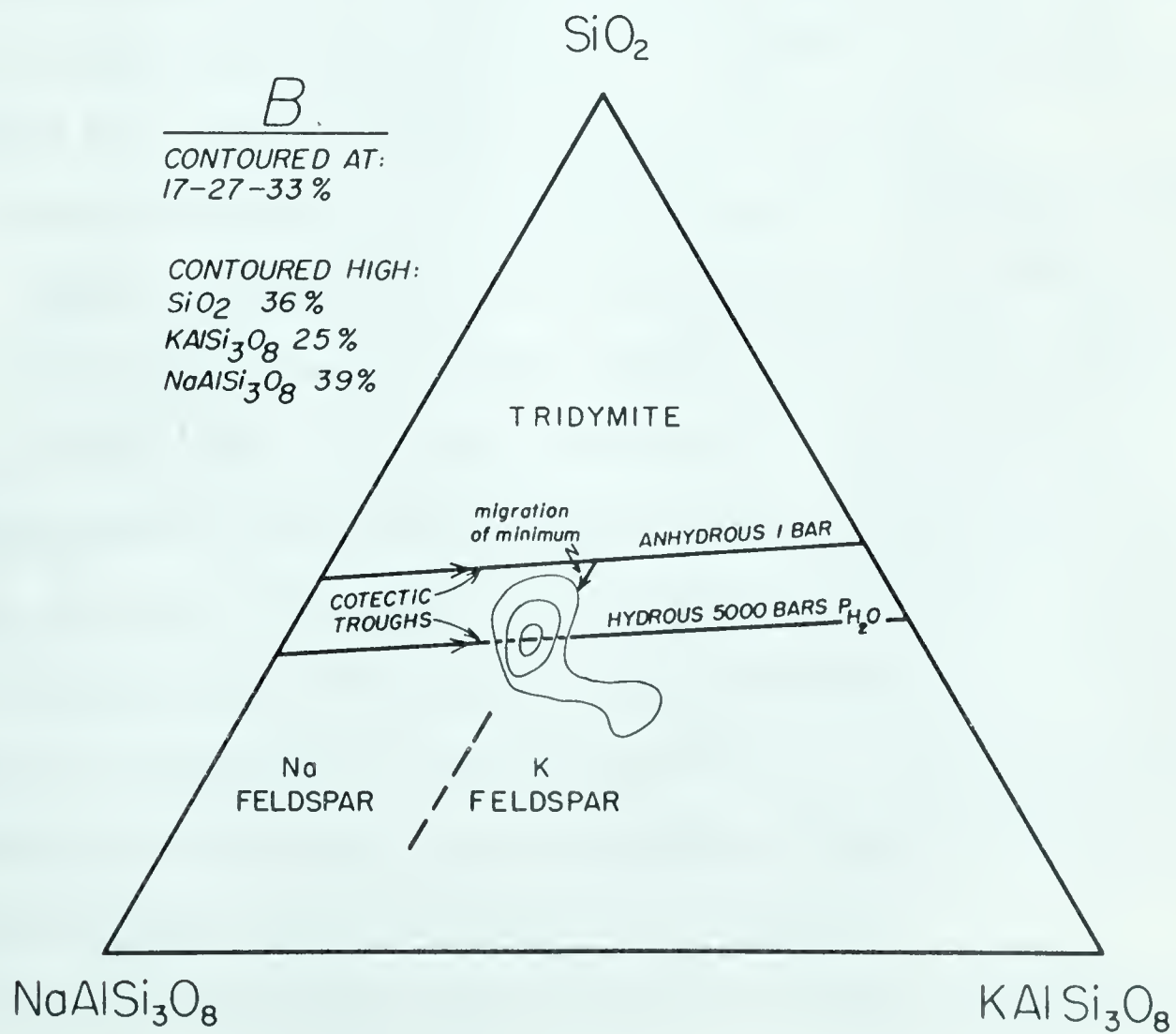
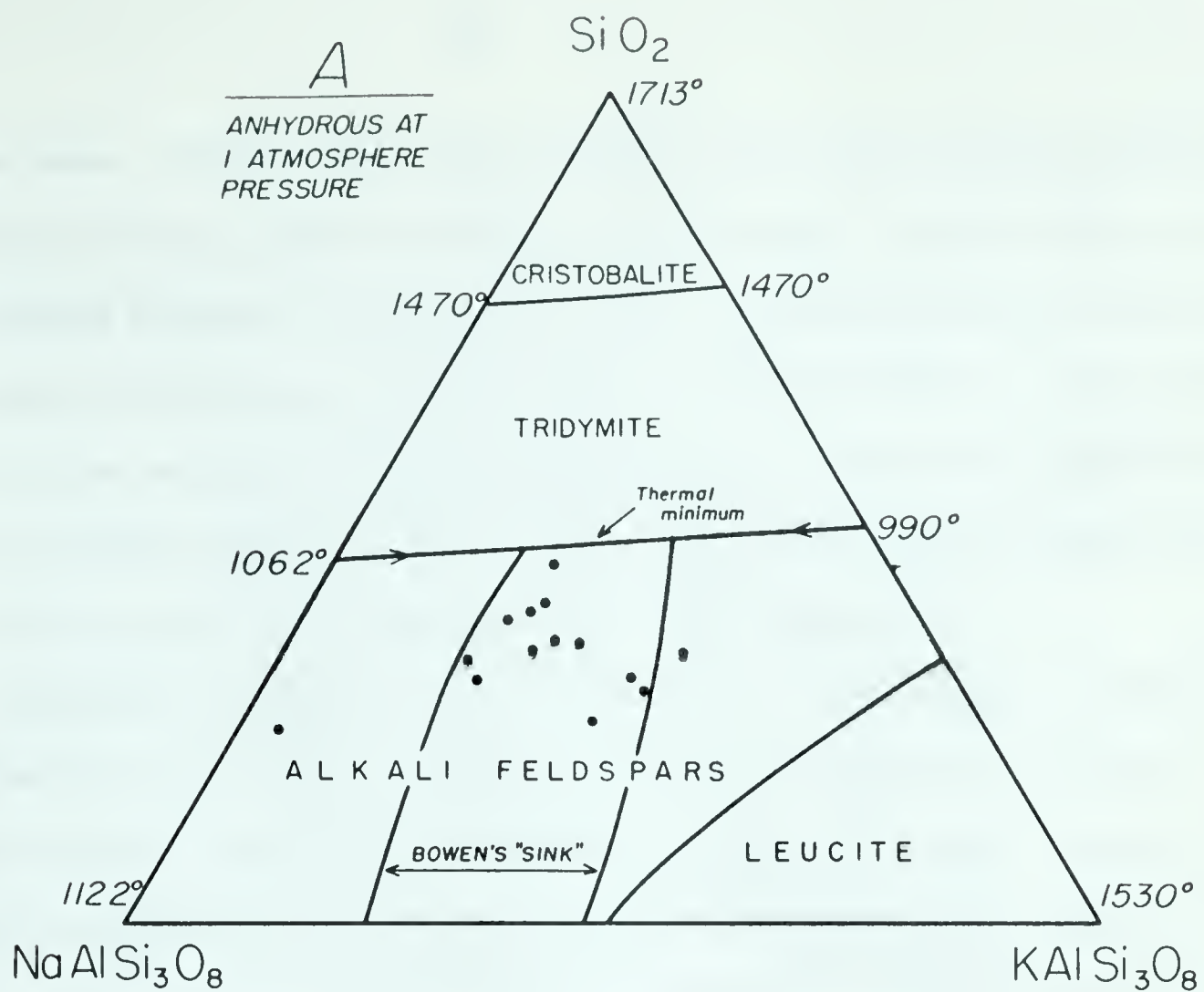


Figure 4. Ternary Plots of Birch Portage Aplites and Pegmatites onto Petrogeny's Residua System

follow (see Bowen, 1937, p. 12). The reasoning for this is that all liquids approaching the plane of the Residua System do so along troughs or valleys in polydimensional space that are directed towards this thermal valley ("sink") in the alkali-alumina silicate plane. All the points within the contours on Figure 4.B lie within this "sink". This trend can be explained for the most part by small scale differentiation, and since the liquid-solidus surfaces of the alkali feldspar melts is very flat, then a small change in temperature causes a large change in the composition of the precipitating crystals.

Petrographic evidence showed that quartz is co-precipitated with feldspars from the margins of the walls up to the central phase where it does not necessarily form a quartz rich zone. Crystallization of feldspar and quartz was therefore cotectic. An examination of diagrams given by Tuttle and Bowen, 1958 (reproduced in Figure 5) shows that with an increase of pressure the minimum on the cotectic trough is moved as shown in Figure 5. Under these conditions the cotectic trough passes through the contoured high on Figure 4.B., and the minimum now lies within the contours.

Petrographic evidence has also indicated subsolvus crystallization of alkali feldspars. Subsolvus crystallization occurs when the melting points of orthoclase and albite are fluxed down to intersect the solvus which occurs at a maximum temperature near 660°C (Tuttle, 1958). The temperature of the solvus surface has been shown to be negligibly affected by high pressures. Recent work at the Geophysical Laboratory, Washington, has shown two methods by which subsolvus crystallization of feldspar occurs. Tuttle and Bowen (1958) showed that under an applied water vapour pressure of 5000 bars the liquidus surface would intersect the solvus. More recently Orville (1963) showed that if the vapour was rich in potassium and sodium chlorides then an applied vapour pressure of only 2,000 bars would be necessary.

Orville was preoccupied with the partitioning, stability relationships and cationic exchange between alkali feldspars and a vapour containing sodium and potassium chlorides in the temperature range 350-700°C and 2000 bars pressure. He

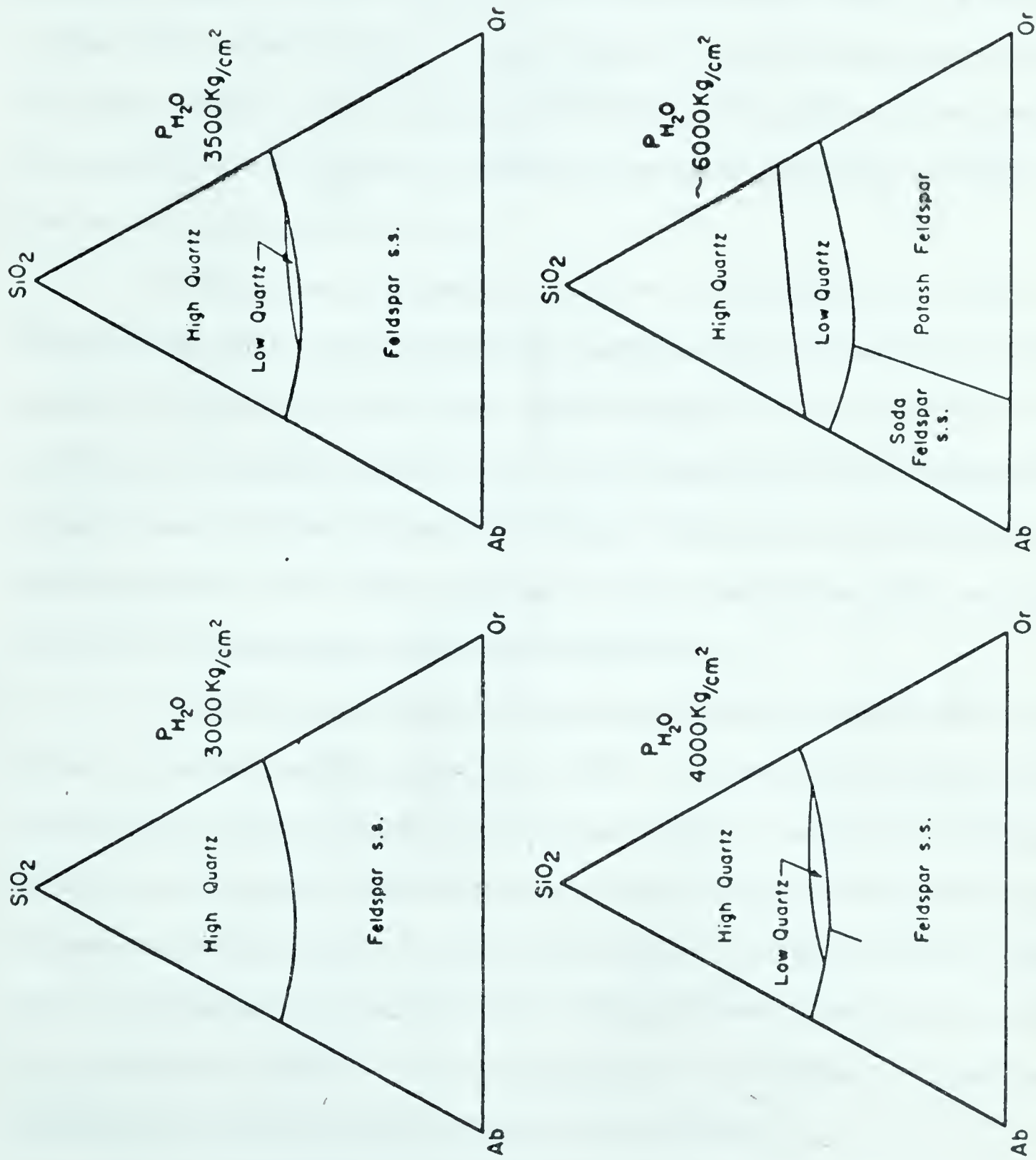


Figure 5. The Effect of Water Vapour Pressure on the Quartz-Feldspar Cotectic Trough (after Tuttle and Bowen, 1958)

shows conclusively that cationic exchange between feldspars and alkali chloride vapours occurs in an attempt to equalize the chemical potentials. In general an alkali chloride vapour with a high $K/K+Na$ ratio will be in disequilibrium with a soda rich feldspar, and the feldspar will become progressively richer in K and the vapour in Na, as the exchange process proceeds. The process would cease when the vapour composition reaches an equilibrium and Orville shows that at a temperature at 600°C and total pressure of 2000 bars the vapour composition equilibrium is reached at $\text{Na}_{77}\text{K}_{23}$ mol per cent.

Orville derives an important conclusion which states that the cationic exchange will occur under disequilibrium conditions which are determined by the variation in composition of the vapour and/or feldspars. The result of this disequilibrium is continued replacement of K-rich feldspars by Na-rich feldspars at higher temperatures and replacement of Na-rich feldspars by K-rich feldspars at lower temperatures. The relation expressed for vapour-solid disequilibria was found by Orville to be similar for melt-crystal disequilibria.

At 2000 bars pressure the exchange occurred quite rapidly at 400°C and was made to go under laboratory conditions at 350°C . The existence of a sodium rich residual melt containing dissolved chlorides would tend to lead to the development of replacement perthites (observed petrographically) but since at low temperatures the tendency is for K-feldspar to replace Na-feldspar the degree of perthitization may be small (observed petrographically). Of significance is the fact the reaction could take place at 350°C . This would facilitate the production of virtually pure orthoclase with only 5% albite remaining in solid solution.

An estimation of the pressure of formation can thus be gained by these experimental studies. Conditions expressed by Orville (1963) probably do not obtain under normal magmatic crystallization, but the presence of dissolved salts is clearly demonstrated by fluid inclusions in beryl (e.g. Cameron, 1953). Thus a minimum

pressure of perhaps 3,000 bars is necessary. This corresponds to a minimum depth of burial of 36,000 feet assuming 1 bar = 12' sediment.

Accepting the temperatures on the solvus surface of Tuttle and Bowen's diagrams (1958) as correct, it is possible to estimate the temperature of formation from the chemical data. Thus the temperature of initial crystallization of the contoured high given in Figure 4.B., is a maximum of 650°C. Low quartz is almost universal in pegmatites and this is formed as a primary mineral at 575°C. Tuttle and Bowen (1958) show however that this inversion point is raised to the temperature of the solvus maximum by a water vapour pressure of 3,000 bars.

The temperature of formation of the central phase feldspars was at a maximum of 560°C and it seems probable that a minimum salt-water vapour pressure of about 4,000 bars is needed in this case. This corresponds to a burial depth of 48,000 feet or about 9 miles and this is a reasonable depth in view of the metamorphism of the country rocks.

It seems probable therefore that crystallization of the granitic melts began at a salt-water vapour pressure near 4,000 bars and a temperature near 650°C. At the time of crystallization of the central phases the temperature had probably decreased to about 560°C. The system probably became all solid at a temperature not less than 550°C, thereafter solid state diffusion occurred involving a re-ordering of the Si-Al atoms in the feldspars. The net result of this was the exsolution of albite from orthoclase. By means of the method of Orville, (1957) ($\Delta^{\circ}2\theta$, KBrO_3 101 minus K-Fsp. $\bar{2}01$) it was found that only about 5% $\text{NaAlSi}_3\text{O}_8$ still remains in solid solution with KAlSi_3O_8 and composition of the exsolved albite is pure. At a temperature of 500°C Bowen (1950) showed that there would be about 22% albite still in solid solution with orthoclase. Unfortunately, no experimental data is available for solvus curve below 500°C principally due to the difficulty of very slow reaction times under laboratory conditions. Extrapolation of the curve as it exists at present suggests that continuous exsolution of albite

from orthoclase would leave the orthoclase 95% pure at a final adjustment temperature of 200°C. This figure seems a little low, but it is interesting to note that Lambert (personal communications) working on similar relationships for feldspars in the Glen Dessary System Complex, Scotland, concluded final adjustments within the feldspar took place at a temperature less than 400°C. Perhaps a figure of 300°C is more realistic and Smith (1960) suggested that the inversion of orthoclase to microcline begins at about 300°C. If this value is correct then final adjustment probably occurred at a temperature somewhere near 300°C. as central phase K-feldspars have a 75% triclinicity.

Classification of the Pegmatites and Aplites

On the basis of experimental studies Tuttle and Bowen (1958) classified granitic rocks into Hypersolvus and Subsolvus granites, Syenites and Nepheline syenites. The subsolvus granites were divided into three main types each with three further subdivisions (see Tuttle and Bowen, 1958, page 130). On the basis of this classification, the Birch Portage granitic rocks belong to the group, microcline - low albite - perthite subsolvus granite with normative albite exceeding 30%. The contoured high on Figure 4.B. has the composition of 36% quartz, 25% orthoclase, and 39% albite.

PLATE 1

A. Porphyroblastic Hornblende Plagioclase Gneiss. D.R.550-63.

Slightly twinned porphyroblastic albite displacing groundmass, with 2 smaller (white) porphyroblasts of hornblende in a quartz-plagioclase (andesine) biotite groundmass. (Crossed nicols. X12)

B. Hornblende K-feldspar Gneiss. D.R.527-63. Hornblende

porphyroblast and sphene with small biotite flakes in a quartz-feldspar groundmass. (Plane polarized light. X30)

C. K-feldspar Gneiss (Unit 4a). D.R.599-63. Typical subhedral

granular texture of the gneiss with occasional twinned oligoclase, many untwinned grains of oligoclase, much strained quartz, K-feldspar and little biotite. (Crossed nicols. X30)

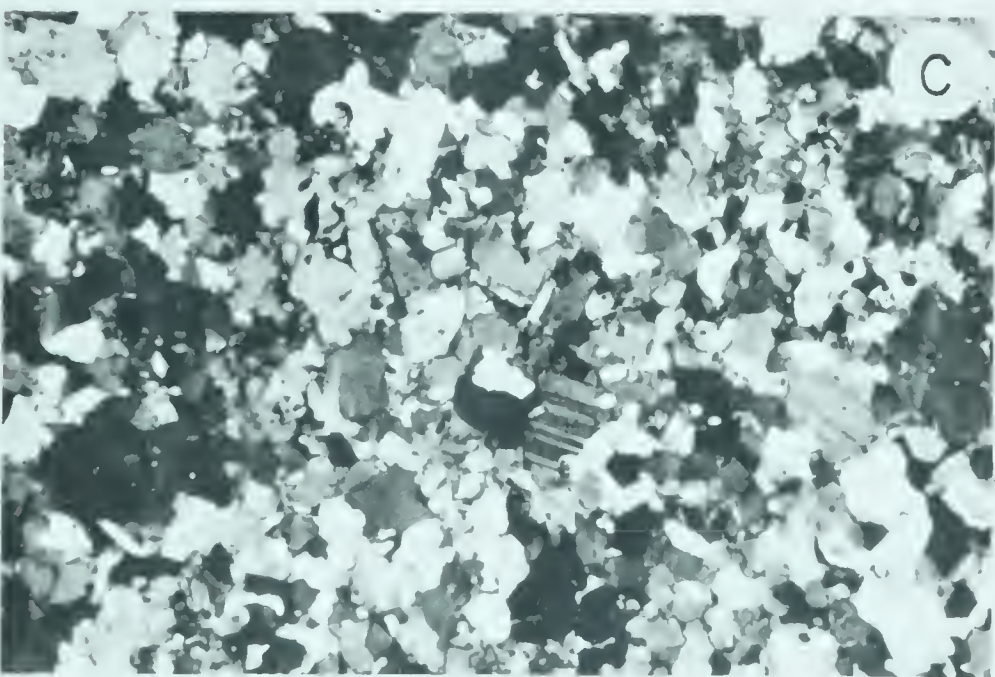
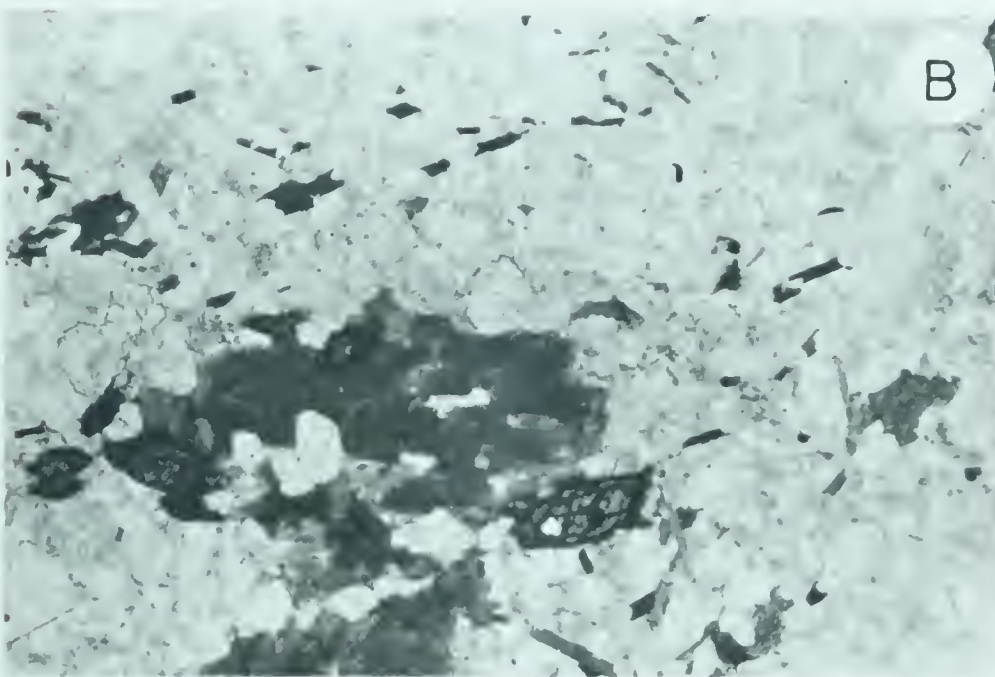
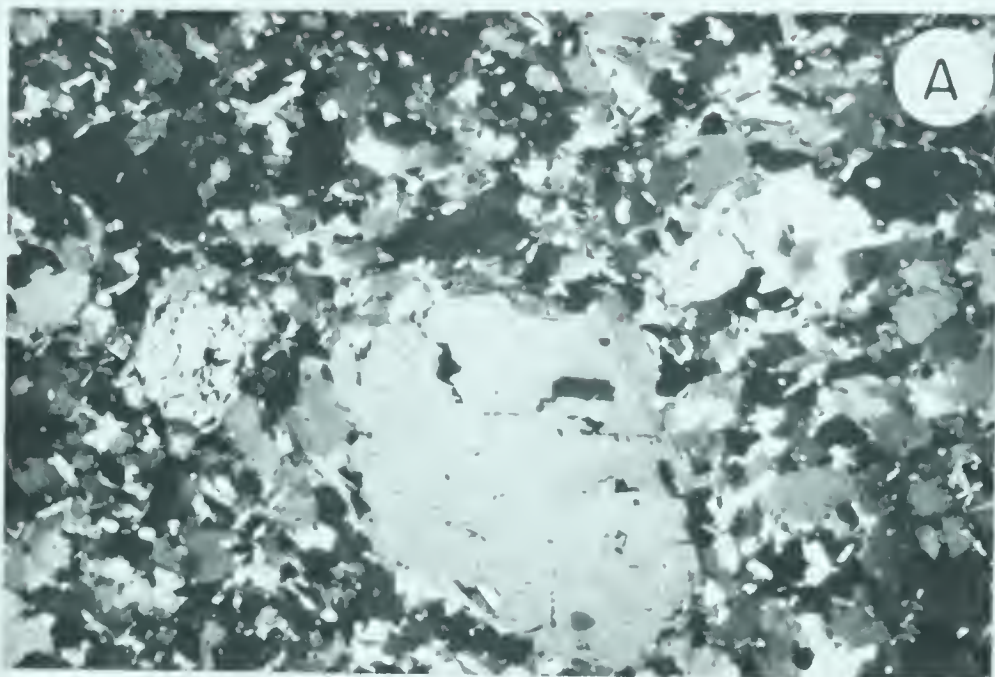


Plate 1. Photomicrographs of Gneisses

PLATE 2

A. Plagioclase Amphibole Gneiss. D.R.598-63. Femic layer in banded gneiss showing large hornblende porphyroblast enclosing quartz and feldspar, small subaligned biotite flakes in a quartz-feldspar (calcic andesine) groundmass. (Plane polarized light. X30).

B. Metamorphosed basic dyke. D.R.530-63. Some dark prominent and irregular chlorites (penninite) and tremolite altering to chlorite in a highly altered plagioclase matrix. (Plane polarized light. X75)

C. Metamorphosed basic dyke. D.R.513-63. Different texture from B. (above). Small hornblende porphyroblasts in an interlocking mass of biotite, plagioclase and magnetite. (Plane polarized light. X30)

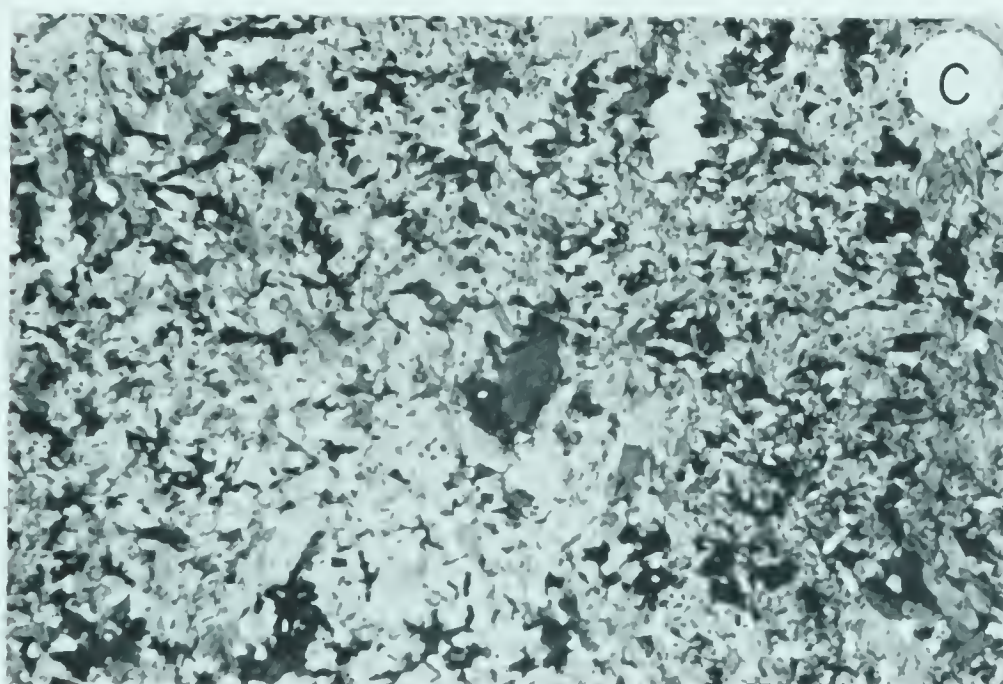
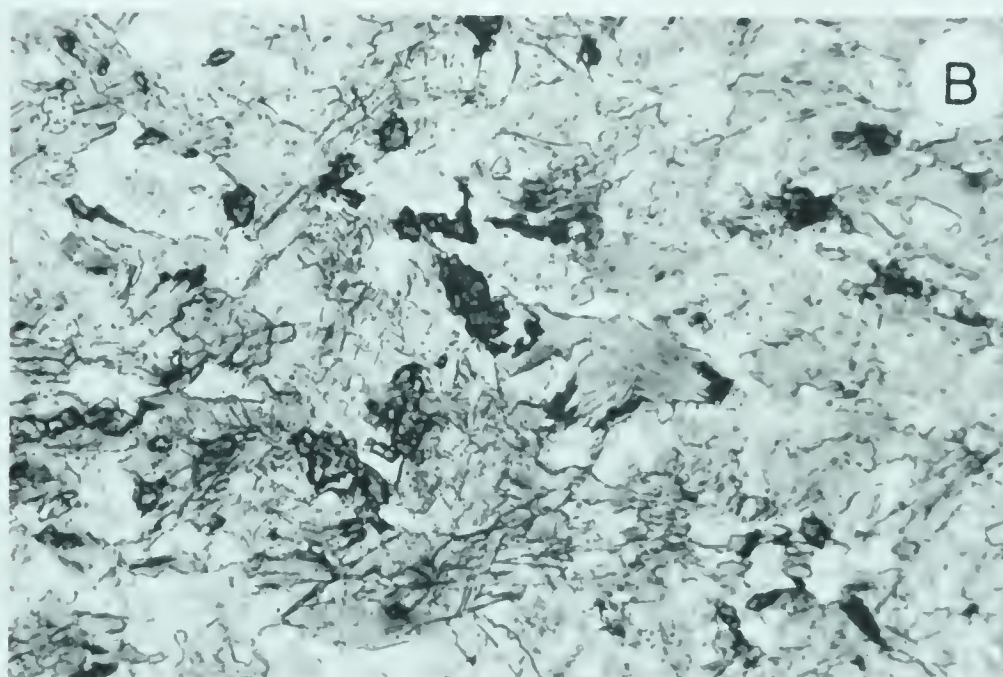


Plate 2. Photomicrographs of Gneisses and Basic Dykes

PLATE 3

A. Red Aplite D.R.510-63. K-feldspars phenocryst (stained) showing development of "cross-hatched" twinning (?) and phenocrysts of albite showing combined carlsbad and albite twinning, in a matrix of twinned albites, quartz and little magnetite. (Crossed nicols. X30)

B. Red aplite. D.R.533-63. K-feldspar phenocryst in the sugary-textured quartz-feldspar groundmass. (Crossed nicols. X30)

C. Cream Aplite. D.R.519-63. Phenocrysts of untwinned K-feldspar in a matrix of fine quartz and albite, some of the large albites showing "cross-hatched" twinning, probably as a result of strain (glide twinning). (Crossed nicols. X12)

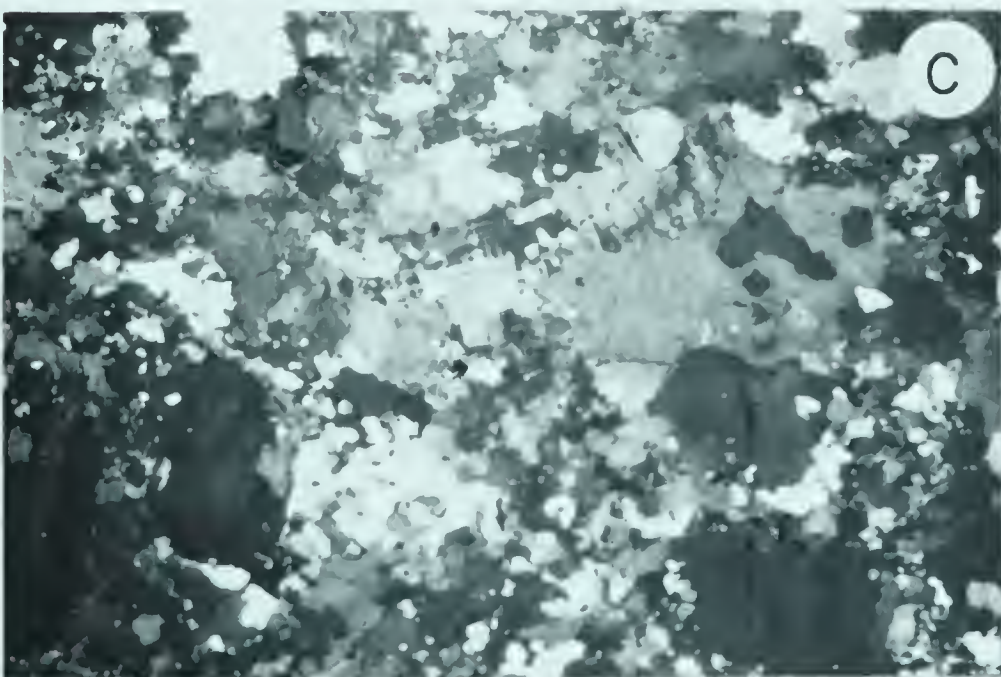
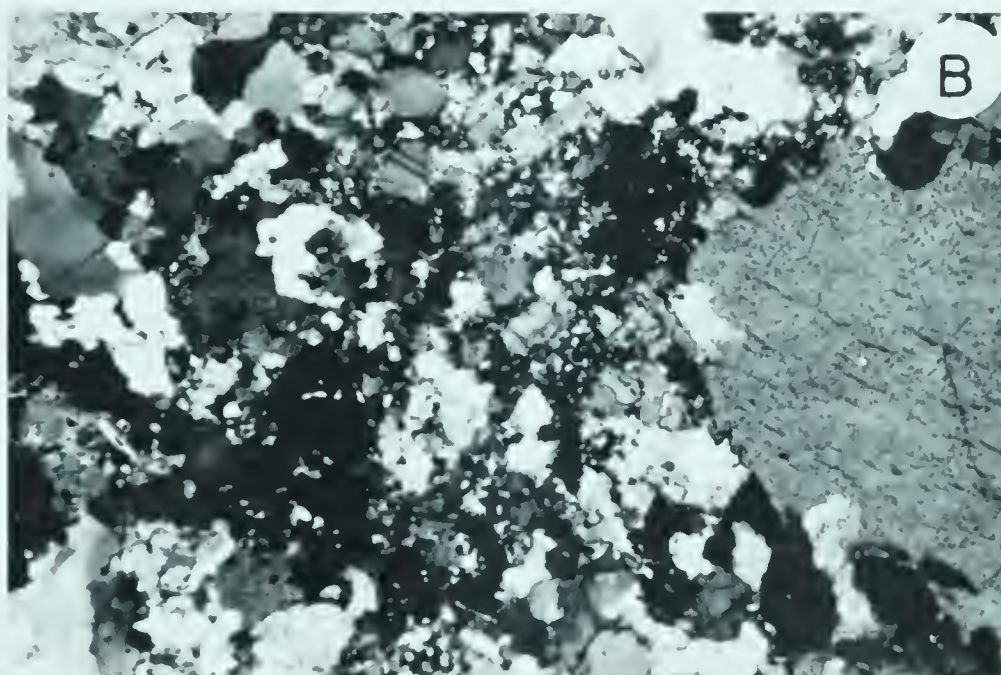
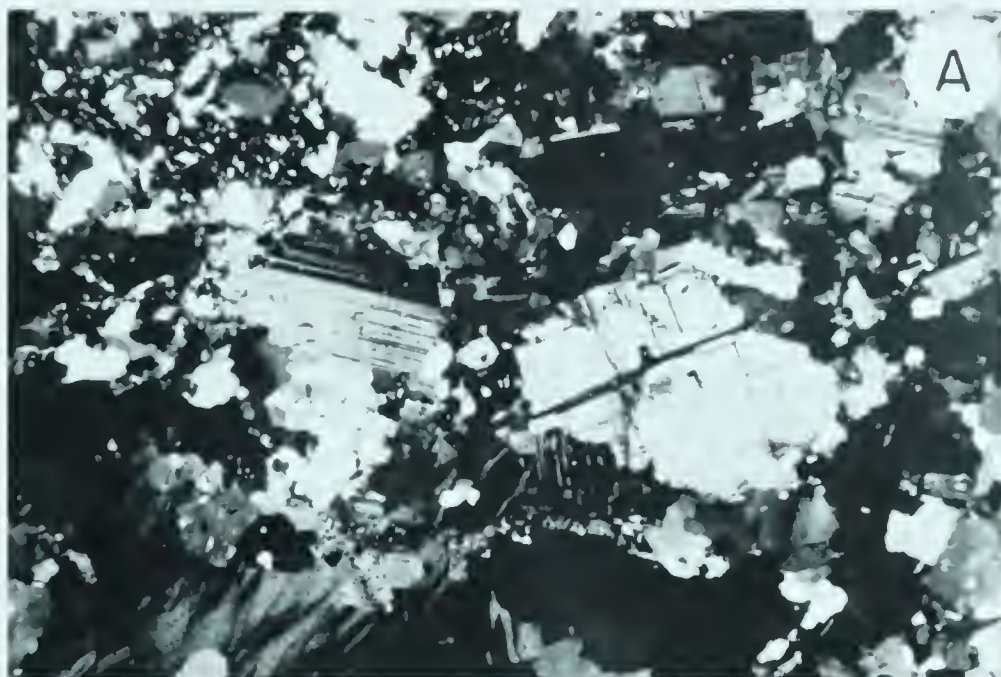


Plate 3. Photomicrographs of Aplites

PLATE 4

A. Two phase Dyke (Pegmatite and Aplite). D.R.526-63. Illustrating two large microcline grains one of which is partly replaced by quartz lenses in optical continuity. (Crossed nicols. X12)

B. Barren Pegmatite. D.R.539-63. Disintegrated xenolith in pegmatite showing partial assimilation of magnetite and production of biotite. (Plane polarized light. X12)

C. Barren Pegmatite. D.R.556-63. Twin gliding in albite. (Crossed nicols. X75)

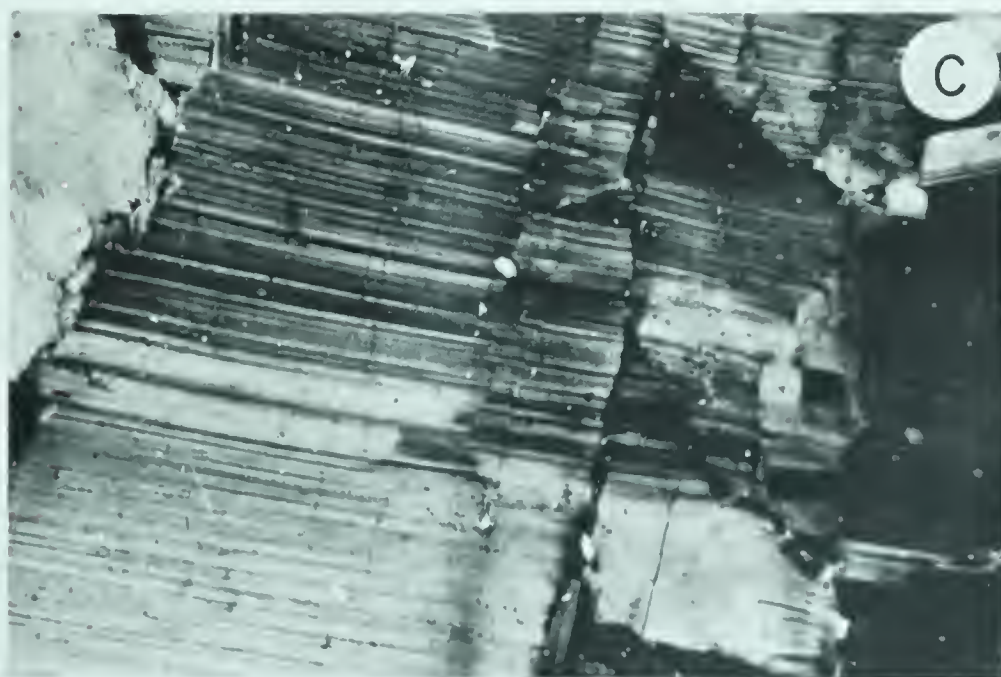
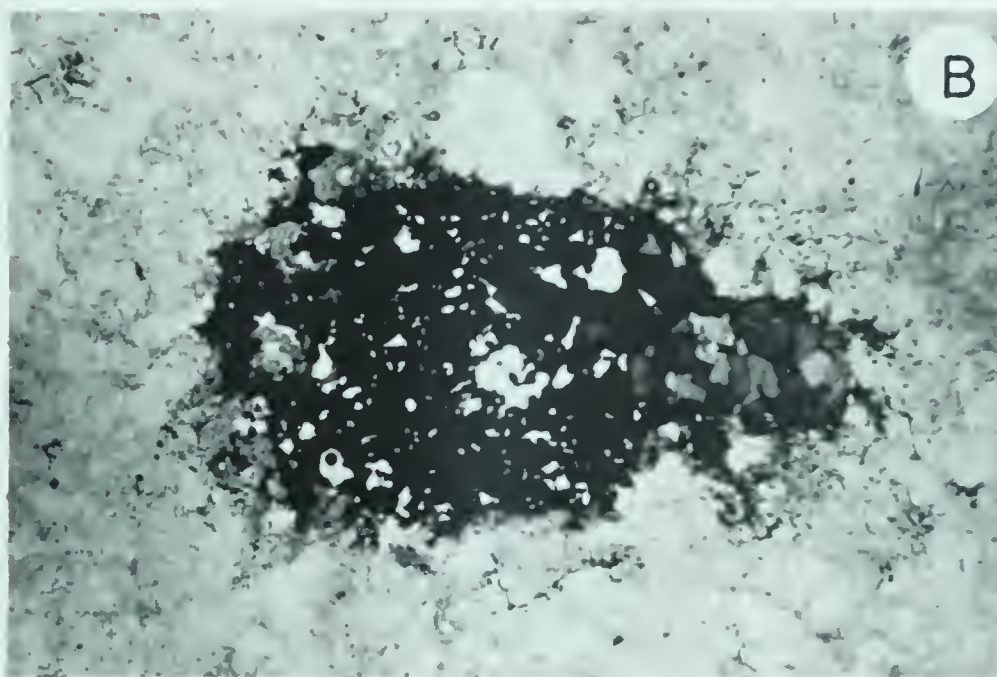
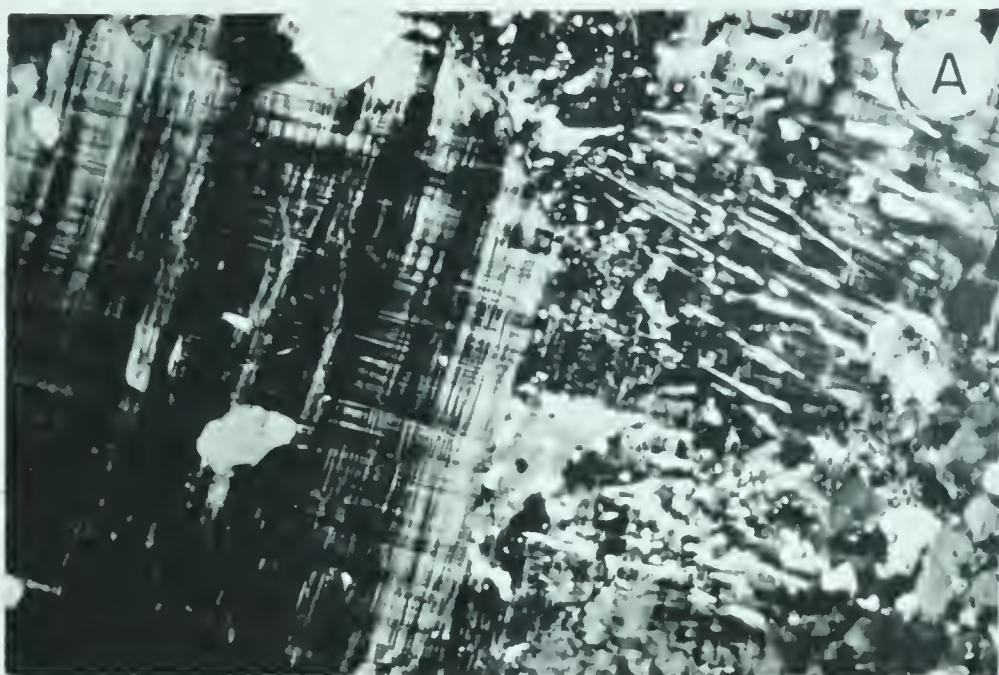


Plate 4. Photomicrographs of Barren Pegmatites

PLATE 5

A. Beryl pegmatite. D.R.601-63. Strained suture textured quartz in finer phase of the beryl pegmatite dyke. (Crossed nicols. X12)

B. Beryl pegmatite. D.R.521-63. Primary precipitate spessartite garnets enclosed in a large plate of K-feldspar, with rounded structureless quartz. (Plane polarized light. X12)

C. Yellow beryl. D.R.522-63 (001) section of beryl showing a fluid inclusion which exhibited very strong second order birefringence colours. (Crossed nicols. X480)

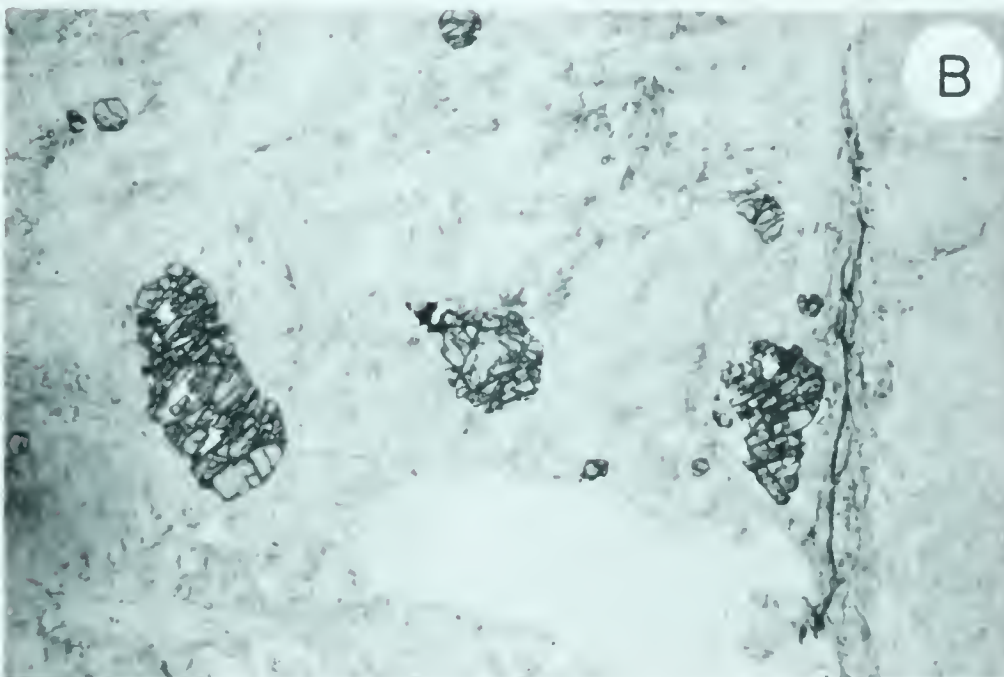


Plate 5. Photomicrographs of Beryl Pegmatites

PLATE 6

A. Perthite. D.R.608-63. Lenticle of twinned albite exsolved from untwinned K-feldspar. (010) albite twinning, perpendicular to well developed (001) K-feldspar cleavage. (010) cleavage poorly developed and section probably h0l. (Crossed nicols. X120)

B. Perthite. D.R.608-63. Smaller scale photomicrograph of A (above) showing the twinned albite exsolution lamellae irregularly continued by replacement perthitization. (Crossed nicols. X12)

C. Perthite D.R.601A-63. Fine untwinned exsolution lamellae of albite in K-feldspar showing a well developed (001) cleavage. Angular intersection (001) cleavage and exsolution lamellae $(60\bar{1}) = 71^\circ$ (theoretically 73°). Thus, this is an almost perfect (010) section [(010) parallel with plane of photograph]. Note diffraction of cleavage by exsolution lamellae. (Crossed nicols. X190)

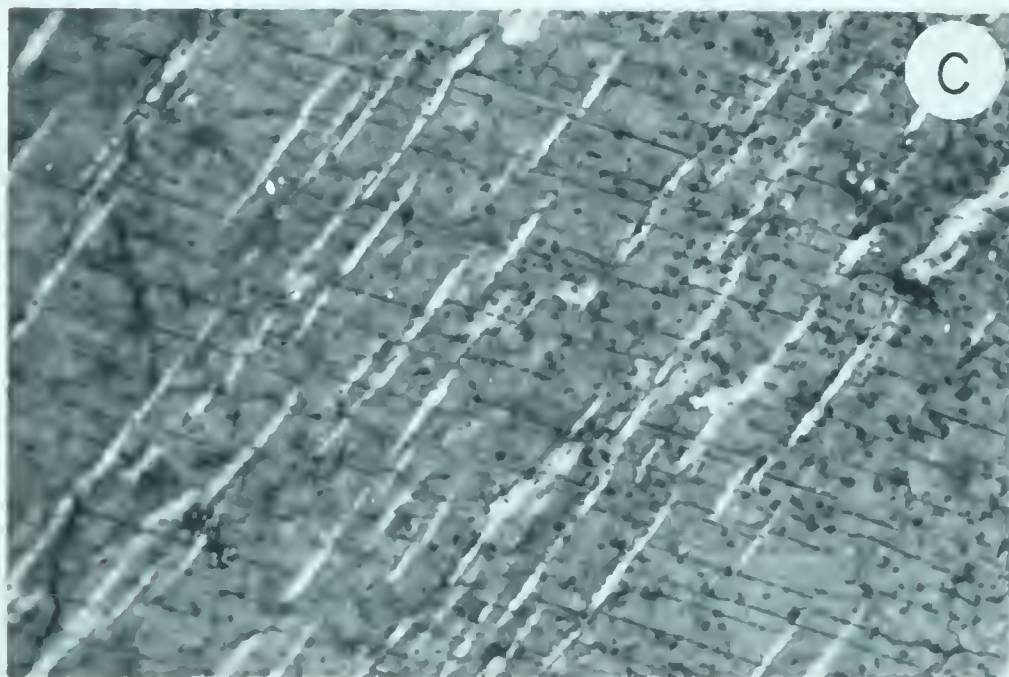
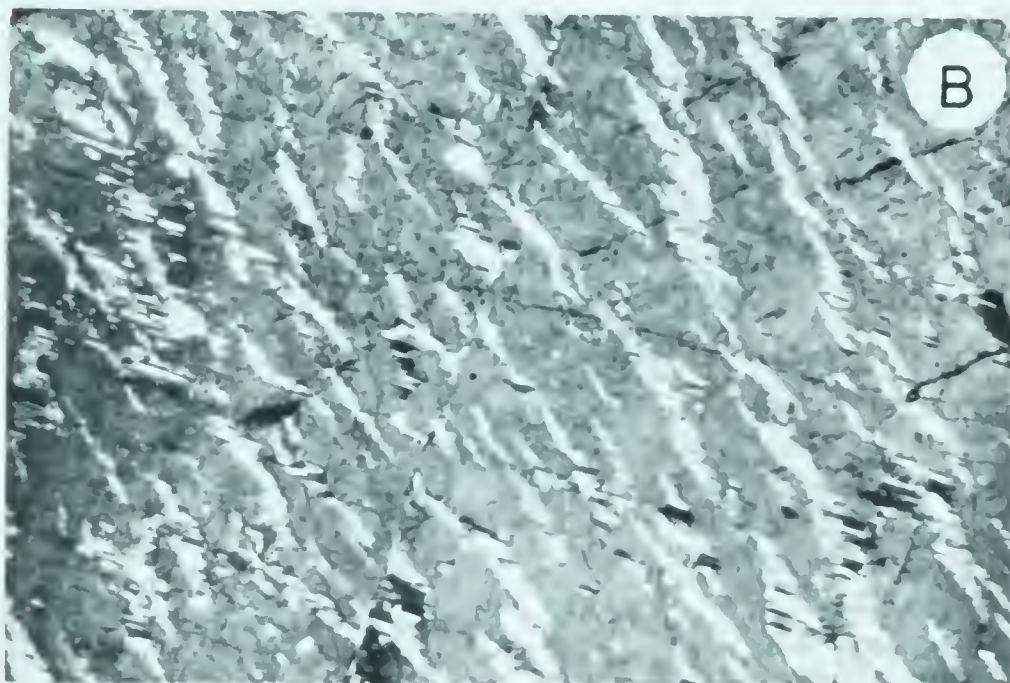
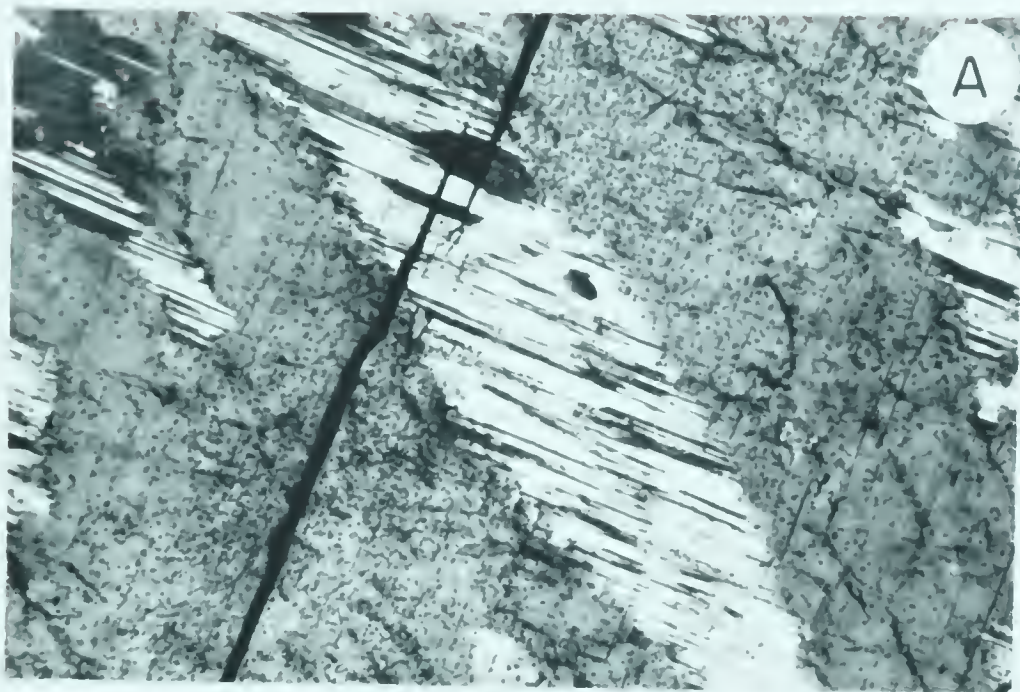


Plate 6. Photomicrographs of Perthites

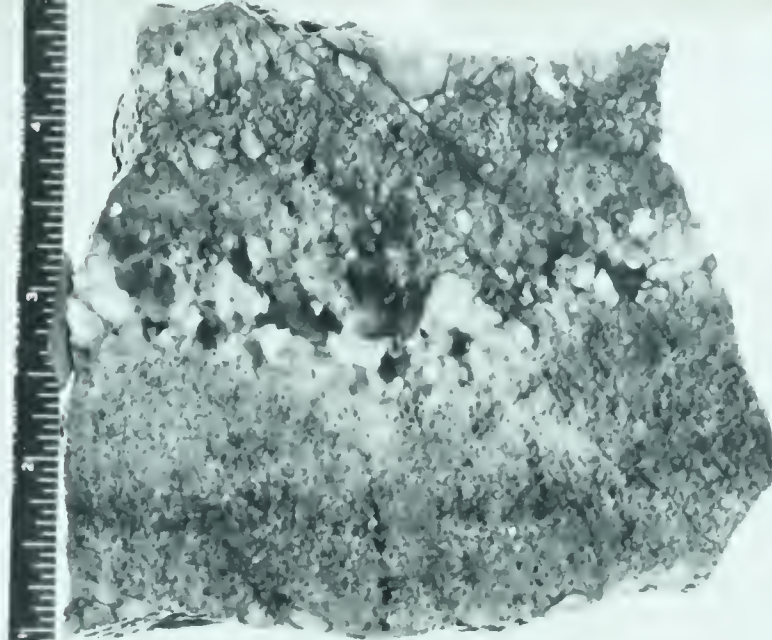
PLATE 7

A. Beryl Pegmatite. D.R.601A-63. Two phase dyke illustrating the rhythmic banding of coarse and fine phases. Note hexagonal beryl crystal in central phase.

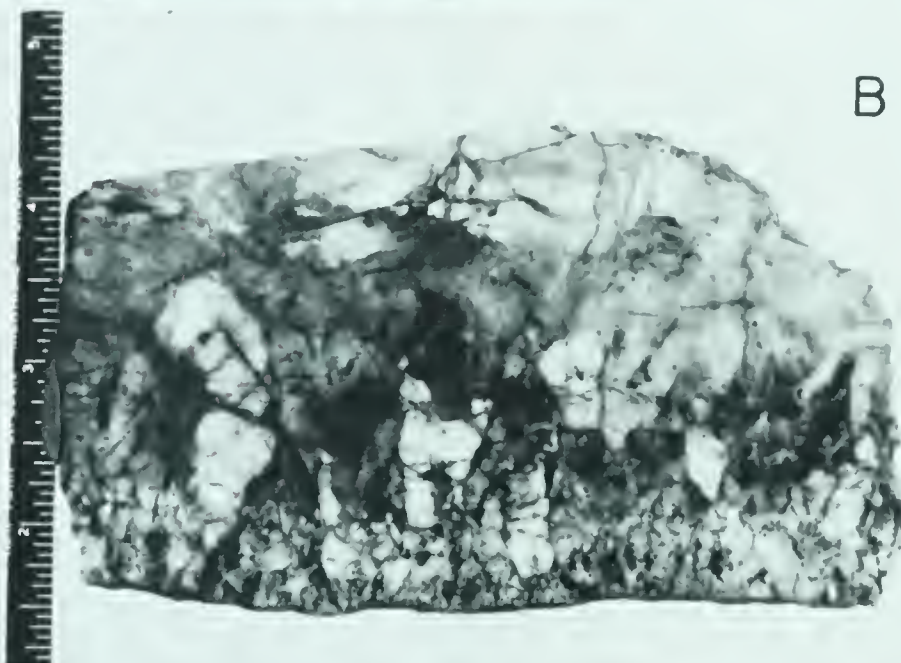
B. Beryl pegmatite. D.R.549-63. Illustrating parallelism of crystal growth from the margins of the dyke and non-parallel, but rather planar, growth in the central phase. Note narrow chilled margin.

C. Barren pegmatite D.R.556-63 showing parallelism of crystal growth perpendicular to the dyke wall and the co-precipitation of quartz, albite and K-feldspar. Note narrow chilled margin.

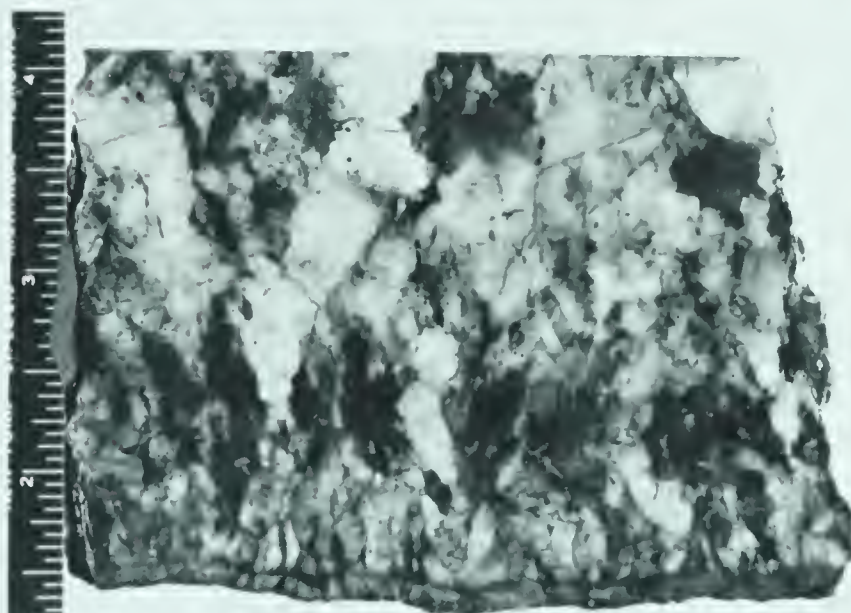
D. Barren pegmatite. D.R.588-63. Indicating a wide "chilled" margin and co-precipitation of albite, K-feldspar, and quartz. Note dark altered xenoliths consisting of a biotite-hematite-garnet aggregate.



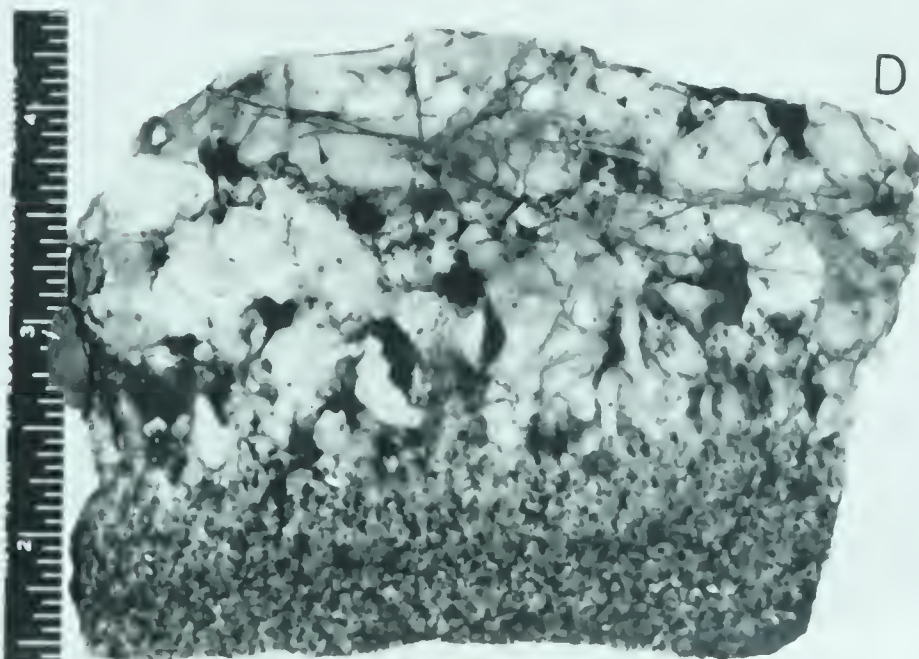
B



C



D



CENTRAL
PHASE

↑
← DYKE WALL →

Plate 7.

Structure of Pegmatites

CHAPTER 4

SOME PHYSICAL AND CHEMICAL ASPECTS OF BERYL

The physical and chemical techniques employed in this study are described in Appendix I.

Crystallography

Beryl belongs to the holosymmetric class of the hexagonal system, being dihexagonal bipyramidal. Its point group is $\frac{6}{m} \frac{2}{m} \frac{2}{m}$. This is a hexad axis at the intersection of two sets of three vertical planes of symmetry, two sets of three diad axes normal to these planes, a plane of symmetry normal to the hexad axis, and a centre of symmetry. The dihexagonal character of beryl and the consequent 2 sets of three planes and diad axes, arise by the superposition of the (110) prism onto the (100) prism. The (110) prism is the least dominant and does not always occur e.g. Birch Portage beryls.

Crystal Structure

The crystal structure of beryl is relatively simple and has two essential characteristics:

- (1) In the (001) plane the structure can be seen to consist of hexagonal rings of six Si-O tetrahedra which envelop a channel that is continuous down the length of structure parallel to 'c'. According to Deer (1962) no atom is nearer than $2.55 \text{ \AA}$ to the centers of these channels which forms hexad axes. Within the Si-O tetrahedral rings two of the oxygen atoms in each SiO_4 group are shared by the SiO_4 groups on either side thus giving a ratio of Si_6O_{18} .

- (2) In any plane perpendicular to (001), the silicon atoms and shared oxygens can be seen to lie on the same plane of components $0, C/2, C$, thus the first reflection of the (001) form is 002. In between these planes are a single plane of cations containing Al and Be atoms, each Al being coordinated with an octahedral group of six oxygen atoms and each Be atom surrounded by four oxygen atoms in a distorted tetrahedron. In these positions they link the oxygen atoms of the Si_6O_{18} rings both laterally and vertically.

Beryl was structurally analyzed by Bragg and West (1926) who gave the unit cell dimensions as: $a_0 = 9.21 \pm 0.01 \text{ \AA}$, $c_0 = 9.17 \text{ \AA} \pm 0.01 \text{ \AA}$. They gave the cell content as $2 \cdot \text{Be}_3\text{Al}_2\text{Si}_6\text{O}_{18}$, and determined the density by X-ray as 2.661. They derived the space group as $C 6/m \text{ cc}$.

The C centred cell refers to the complete hexagon and not the 120° right prism, this latter being $1/3$ of the hexagonal cell with a P cell and cell content of 2. This 120° right prism contains the full symmetry elements and the cell should be correctly referred to a P. However the double 'c' gliding, places atoms in the centres of the sides of the 120° right prism and for this reason first order X-ray reflections might not be expected. Certainly (001) is cancelled and (002) is the first 'c' reflection. However in the 'c' plane incomplete cancellation of 1st order reflection occurs due to the existence of numerous open channels in the structure which are parallel with the 'c' axis. Thus the sequence of atoms in the (100) and (200) planes when traced across the lattice are as follows:

(100) plane: 5 atoms - 1 space - 5 atoms - 1 space, etc.

(200) plane: 2 atoms - 1 space - 2 atoms - 1 space, etc.

For this reason (100) is the first and strongest reflection appearing on the X-ray pattern. Since the unit cell is a 120° right prism conversion of (100) reflections into a_0 is given by Buerger (1958) as:

$$\frac{1}{d^2_{hkl}} = \frac{4}{3} \frac{(h^2 + hk + k^2)}{a_o^2} + \frac{l^2}{c_o^2}$$

Derivation of c_o from (002) reflections is by a simple multiplication.

Indexing of Beryl

One of the main problems of indexing and consequently of determining accurate unit cell parameters in the fact that in beryl the a_o and c_o dimensions are virtually the same, e.g. Deer (1962) accepts Norrish's (1950) data of: $a_o = 9.188$, $c_o = 9.189$. For this reason many combination peaks occur on the pattern and there are often several hkl planes which satisfy the space group for any one d spacing.

The realization of the existence of combination peaks is of prime importance in determining accurate cell parameters. This fact probably explains many of the variations expressed in the literature, e.g. Schaller (1962) derives his parameters from 002 and then uses this value of c_o in reflection of the type (112) (211), (212) etc.; no mention is made of the fact that 002 is coincident with 110. Thus if the obtained c_o is then used to calculate a_o in reflections of the type 112 etc., the answer for a_o (and of c_o) must necessarily be incorrect.

In the writer's opinion only "clean" reflections should be used in the determination of cell parameters and a careful investigation shows that only the following reflections are "clean" and can be read easily:-

(100) odd orders only, as even reflections are coincident with (102) forms

(112) to determine c_o knowing a_o

Clearly if one uses all h00 then the even orders will give a biased answer.

The cell parameters determined by the writer are as follows:

TABLE 13

	<u>a_o ($\pm$ 0.001)</u>	<u>c_o ($\pm$ 0.001)</u>
609 aquamarine	9.2121 Å	9.1853 Å
522 yellow	9.2006 Å	9.1868 Å
551 green	9.2545 Å	9.1892 Å

The fairly constant value of c_o is encouraging and it will be shown in the section, "Crystal Chemistry", that the variation of a_o is due mainly to inclusion of alkali elements. These values are in reasonable agreement with those published in the literature, e.g. see Table 14, below:

TABLE 14

	$\underline{a_o}$	$\underline{c_o}$
Bragg (1926)	9.21	9.17
Norrish (1950)	9.186	9.189
Huttenlocher (1954)	9.22	9.18
Sosedko (1957)	9.221	9.202
Sosedko (1957)	9.221	9.228
Sosedko (1957)	9.219	9.246
Schaller (1962)	9.30	9.20

An example of an indexed pattern is given in Table 15 (D.R. 522).

Specific Gravities and Cell Parameters

Specific gravity can be measured two ways, either directly or from the analyzed chemical composition and unit cell dimensions according to the equation:

$$S.G. = \frac{M.W. \times Z}{V \times A_v.No.}$$

Where $M.W.$ = Molecular Weight, Z = no. of formula units in a unit cell, V = Volume of unit cell in cubic Angstrom units or 10^{-24} cm^3 and $A_v.No.$ = Avogadro's Number, used to convert atomic mass units into grams.

Theoretically both specific gravities should be identical and practically they usually are within experimental error. If they are not, this is often an indication of the inaccuracy of unit cell dimension measurements. The author found that in the case of beryl the calculated specific gravity was lower than the measured specific gravity. Thus:

TABLE 15
X-Ray Diffraction Data for Yellow Opaque Beryl

<u>2θ</u>	<u>dÅ[°]</u>	<u>hkl</u>	<u>I</u>
11.13	7.9428	100	100
19.38	4.5762	002, 110	31
22.33	3.8095	200, 102	32
27.41	3.2511	112	62
29.64	3.0113	202, 210	19
31.22	2.8624	211, 121	50
33.65	2.6611	300	3
35.62	2.5183	212, ?	16
39.12	2.3007	400	7
40.80	2.2097	130, 104	7
42.00	2.1493	(213) 311	11
43.97	2.0575	222, 114	3
45.55	1.9897	400, 204	16
49.77	1.8305	230, 214	2
50.86	1.7937	4 $\bar{1}$ 1?	14
52.62	1.7378	140	14
53.53	1.7104	322, 411	9
56.55	1.6261	224	12
57.77	1.5946	500	5
58.75	1.5730	215, 702	3
60.38	1.5315	006, 330	2
61.18	1.5136	404	8
64.05	1.4505	116, 332, 6 $\bar{2}$ 1	10
65.08	1.4320	422	9

TABLE 15 (continued)

<u>2θ</u>	<u>d$\text{\AA}$^o</u>	<u>hkl</u>	<u>I</u>
68.50	1.3686	216	2
70.95	1.3273	306,600	1
74.18	1.2772	250,226,602	7
75.00	1.2653	251,424	7
79.42	1.2056	217	4
81.60	1.1768	523	1
83.97	1.1515	440	1
84.18	1.1491	008	2
85.10	1.1390	700(108)	1
87.26	1.1163	118,442,208	10
90.51	1.0845	336	1
91.53	1.0750	426,218,622	1
92.35	1.0676	353,345	1
94.46	1.0493	156	1
94.62	1.0479	444,228	1
100.90	0.9989	630,606,451	1
101.12	0.9974	408	1
101.50	0.9947	800	1
111.24	0.9269	550	1
113.53	0.9209	00.10	1
114.53	0.9158	460,428	2
117.48	0.9011	522	1
119.08	0.8936	813,455,8 $\bar{2}2$,900	2
128.71	0.8542	22.10	1

ANALYST: Dennis Radcliffe

TABLE 16

	<u>S.G. Measured</u>	<u>S.G. Calculated</u>	<u>Difference</u>
609	2.715	2.686	0.029
522	2.722	2.681	0.041
551	2.791	2.739	0.052

If we assume that mass is correct and the volume incorrect, calculations show that the unit cell parameters have to be reduced to values which are unacceptable. The changes in unit cell parameter of D.R. 522 to account for this discrepancy would be as follows: Assuming c_0 correct which seems likely considering c_0 of the different beryls, then a_0 would have to be changed from $9.200\text{\AA}$ to $9.121\text{\AA}$ a change of $0.079\text{\AA}$. The writer considers the determination of unit cell parameters better than $\pm 0.01\text{\AA}$.

If we assume the volume to be correct then the change of mass required would be equivalent to 2.40% Fe_2O_3 the heaviest oxide, 3.76% Al_2O_3 or 6.38% SiO_2 . These values are outside the range of acceptable experimental errors. An alternate explanation is that the determination of measured specific gravity is inaccurate, but this is unlikely as the method used gives a precision of ± 0.001 .

Three possible reasons for the difference can be suggested:

- (1) Damon (1958) showed that Precambrian beryls contained up to 100 times more excess rare gases that can be accounted for by radioactive decay (e.g. $\text{He}^4\text{Ar}^{40}$), and that CO_2 formed 75% of water free gases. Assuming these gases were not included in the measurement of H_2O^+ , then this might be a contributory factor. However calculations by the author on Damon's data showed that the maximum content of water free gases could only be 30 ppm.

- (2) The open channels in the beryl structure are known to contain alkalis and water. It is possible during the determination of measured specific gravity that water entered the channels "zeolitically" and the method of computation of specific gravity is such that the value of measured specific gravity would be higher. It is shown in the section, "Crystal Chemistry" that these channels were already partially filled by alkalis and water, and even if water could enter the channels quickly and easily in the manner implied, then the contribution would still only be small.
- (3) Water is not usually included in the calculations of the formulae of beryl. However if 1% water is used in the calculation and computed as a separate mineral, the difference in total calculated density would only be 0.001.

Thus, these three suggestions could only be very minor factors affecting the difference in specific gravity.

An obvious conclusion is that the writer made serious errors during experimentation and/or calculation of the specific gravity data. However examination of the literature shows that all authors found this same discrepancy and in the same direction, e.g. Table 17, below:

TABLE 17

	<u>S.G. measured</u>	<u>S.G. calculated</u>	<u>Difference</u>
Norrish (1950)	2.73	2.641	0.099
Sosedko (1957)	2.72	2.62	0.10
Sosedko (1957)	2.75	2.63	0.12
Sosedko (1957)	2.78	2.73	0.05
Schaller (1962)	2.92	2.84	0.08

In most cases the difference obtained by previous investigators is higher than for the present writer. This is possibly due to their inaccurate determination of unit cell parameters. The case of Schaller has already been discussed and in any case his a_0 and c_0 are not given to sufficiently significant figures. These inaccuracies can be exemplified by examining Sosedko's data which is detailed in his account.

Sosedko's unit cell parameter (after converting from KX to Ångstrom Units) and density data are tabulated below:

TABLE 18

	<u>a_0</u>	<u>c_0</u>	<u>S.G. meas.</u>	<u>S.G. calc.</u>	<u>Diff.</u>
Sample 1	9.221	9.202	2.72	2.62	0.10
Sample 2	9.221	9.228	2.75	2.63	0.12
Sample 3	9.219	9.246	2.78	2.73	0.05

Using the method employed in this study to determine unit cell parameters, the following values were obtained:

TABLE 19

	<u>a_0</u>	<u>c_0</u>	<u>S.G. meas.</u>	<u>S.G. calc.</u>	<u>Diff.</u>
Sample 1	9.2370	9.2048	2.72	2.66	0.06
Sample 2	9.2294	9.2322	2.75	2.67	0.07
Sample 3	9.2353	9.2460	2.78	2.79	0.01

Clearly the recommended method of determining a_0 and c_0 gives a superior calculated specific gravity and it is apparent that the difference of measured specific gravity and calculated specific gravity should be used as a guide to ascertain inaccuracies in unit cell dimensional values. It is proposed by the writer that the values obtained for the cell parameters of beryl should only be accepted when the density difference between measured specific gravity and calculated specific gravity is less than 0.05. Unfortunately the reason for these density discrepancies remains a partly unsolved problem.

Crystal Chemistry

The chemical composition of the three beryls are given in Table 21. The calculations of the formulae were made on the basis of O = 18. The formulae are as follows:

TABLE 20

609.	$(\text{Be}_{2.39}\text{Si}_{.11}\text{Al}_{.11}\text{Fe}_{.14})_{2.75}\text{Al}_{2.00}\text{Si}_{6.00}\text{O}_{18}\text{Na}_{.01}$
522.	$(\text{Be}_{2.77}\text{Al}_{.12}\text{Fe}_{.05})_{2.94}\text{Al}_{2.00}(\text{Si}_{5.86}\text{Al}_{.14})_{6.00}\text{O}_{18}\text{Na}_{.04}\text{K}_{.01}$
551.	$(\text{Be}_{2.27}\text{Al}_{.23}\text{Fe}_{.43})_{2.93}\text{Al}_{2.00}\text{Si}_{6.00}\text{O}_{18}\text{Na}_{.05}\text{K}_{.01}$

In two of the three cases the Si tetrahedral positions are completely filled with Si and in the third sample some Al may enter these sites. The 6 coordinated Al positions are in all cases completely filled with Al. The problem of the 4 coordinated Be sites is noteworthy as in all cases there is insufficient Be to completely fill all the available sites. This fact has given rise to much speculative thought, especially among Russian authors, as to whether or not these positions can be filled with alkalis or if the alkalis are located only in the channels. In any case it is most probable that $\text{Al}^{3+}(0.51\text{\AA})$ and perhaps $\text{Fe}^{3+}(0.64\text{\AA})$ could enter the 4 coordinated sites of $\text{Be}^{2+}(0.35\text{\AA})$. The radius ratios of these elements with respect to O^{-2} are:

$$\text{Be}^{2+} = 0.25, \text{Al}^{3+} = 0.36, \text{Fe}^{3+} = 0.46$$

The most probable positions of the alkalis is in the channels of beryl where they can play a small part in the electrical neutrality of the system. Water most definitely occupies these channels as its exit upon heating occurs without any endothermic effects or change in unit cell dimensions. This indicates that water is "zeolitically" held in beryl. However Ginzburg (1955) has shown that the water is not expelled until a temperature of 800–900°C and Damon (1958) has shown that as little as 60% of the gases are expelled by 1200°C and 100% expulsion is not obtained until 1475°C. At this temperature beryl melts incongruently into liquid and phenakite, Be_2SiO_4 (Van Valkenburg, 1957).

TABLE 21

CHEMICAL ANALYSIS OF BERYL

	<u>609</u>	<u>522</u>	<u>551</u>
SiO ₂	66.00	63.66	63.94
TiO ₂	0.03	---	0.01
Al ₂ O ₃	19.34	20.89	20.98
Fe ₂ O ₃	2.04	0.83	2.04
MnO	0.02	0.01	0.02
MgO	---	---	---
CaO	---	0.01	---
Na ₂ O	0.04	0.24	0.32
K ₂ O	0.03	0.05	0.09
BeO	10.75	12.49	10.08
H ₂ O ⁺	0.59	0.74	1.15
H ₂ O ⁻	0.38	0.38	0.47
Others	---	---	0.01
Total	<u>99.22</u>	<u>99.30</u>	<u>99.11</u>
<u>Trace Elements (ppm)</u>			
SrO	7	8	11
Li ₂ O	7	23	4
Rb ₂ O	16	12	98
Cs ₂ O	4	2	30

609. Aquamarine. Mt. Antero, Colorado

522. Yellow Beryl. Birch Portage, Saskatchewan

551. Green Beryl. Birch Portage, Saskatchewan

ANALYST: Dennis Radcliffe

The high temperature needed to expell water from the beryl structure may be explained in one of two ways:

- (1) It plays a small part in the electrical neutrality of the system as the non-stoichiometric quantities of Be in beryl may give rise to a slight residual positive charge when higher valency cations e.g. Al, enter the (divalent) tetrahedral Be sites, and/or the presence of monovalent alkalis especially Li and Na in the channels may give rise to charge asymmetry.
- (2) The presence of alkalis in the channels effectively blocks the exit of water from the structure at low temperatures.

Bakakin (1962) suggests that the R_2O members occupy definite positions in the channels with the HOH molecules situated opposite the Si-O rings and alkalis opposite the Be-Al cation rings. The reason for this is that since the Si atoms lie closest to the centre of the channels the alkalis will thus be midway between the Si-O rings due to mutual repulsion and consequently the HOH group has to be in the Si-O planes.

Perhaps the best guide to the positioning and possible substitutions is obtained by studying the variation of cell parameters. Since the Si-O ring is a strong unit any substitution within the structure will most probably cause an increase in the c_0 direction i.e. the introduction of ions of larger radius than Be into the Be-Al plane can be more easily accommodated by interplanar expansion. A comparison of the c_0 and Al-Fe atoms in tetrahedral coordination in the writer's beryls supports this suggestion:

TABLE 22

	<u>609</u>	<u>522</u>	<u>551</u>
Atoms (Al, Fe)	0.24	0.31	0.66
$c_0 \text{ \AA}$	9.1853	9.1868	9.1892

Under certain very rare conditions a large atom such as cesium may enter the lattice. Sosedko (1957) provides fairly convincing evidence that this can happen. He shows:

TABLE 23

	<u>Cs₂O (%)</u>	<u>c₀ Å</u>
Sample 1	0.27	9.202
Sample 2	0.67	9.228
Sample 3	4.13	9.246

However in general, the larger atoms do not enter the structure proper but merely reside in the channels. In these positions it is most probable that the lattice would be expanded in the a_0 direction due to mutual repulsion of cations. Figure 6 shows this relationship and other expected trends, and it appears from the figure that a threshold value exists at about 2.5 Mol % R_2O when any further R_2O introduction expands the lattice in the suggested manner. It is interesting to note that Schaller (1962) proposes the same general relationship and in beryl from Arizona with unusually high cesium the lattice was expanded in the a_0 direction and consequently the cesium probably resides in the channels. It is probable that Sosedko's cesium beryls crystallized under very different physico-chemical conditions than those of Schaller.

The sequence of changes with increasing R_2O content shown in Figure 6 is not quite perfect and 522 which has the intermediate R_2O content is lower than the two "extremes" in certain respects, especially a_0 . The structural formulae show that some Al has entered the Si-O hexagonal rings and these facts may be related.

The actual nature of the substitution of alkalis in the beryl structure has been debated by various authors. One fact appears certain that substitution takes place for most part, within the Be-Al cationic layer. Folinsbee (1941) suggests the nature of substitution is: $3 \text{ Alkalis}^{1+} + 1 \text{ Al}^{3+} = 3 \text{ Be}^{2+}$. Deer (1962) modified this into: $2 \text{ Alkalis}^{1+} = 1 \text{ Be}^{2+}$. Another relationship may be suggested: $1 \text{ Alkali}^{1+} + 1 (\text{Al, Fe})^{3+} = 2 \text{ Be}^{2+}$ whereby the trivalent ions enter the beryllium sites whereas the alkalis may simply reside in the open channels or enter the lattice proper.

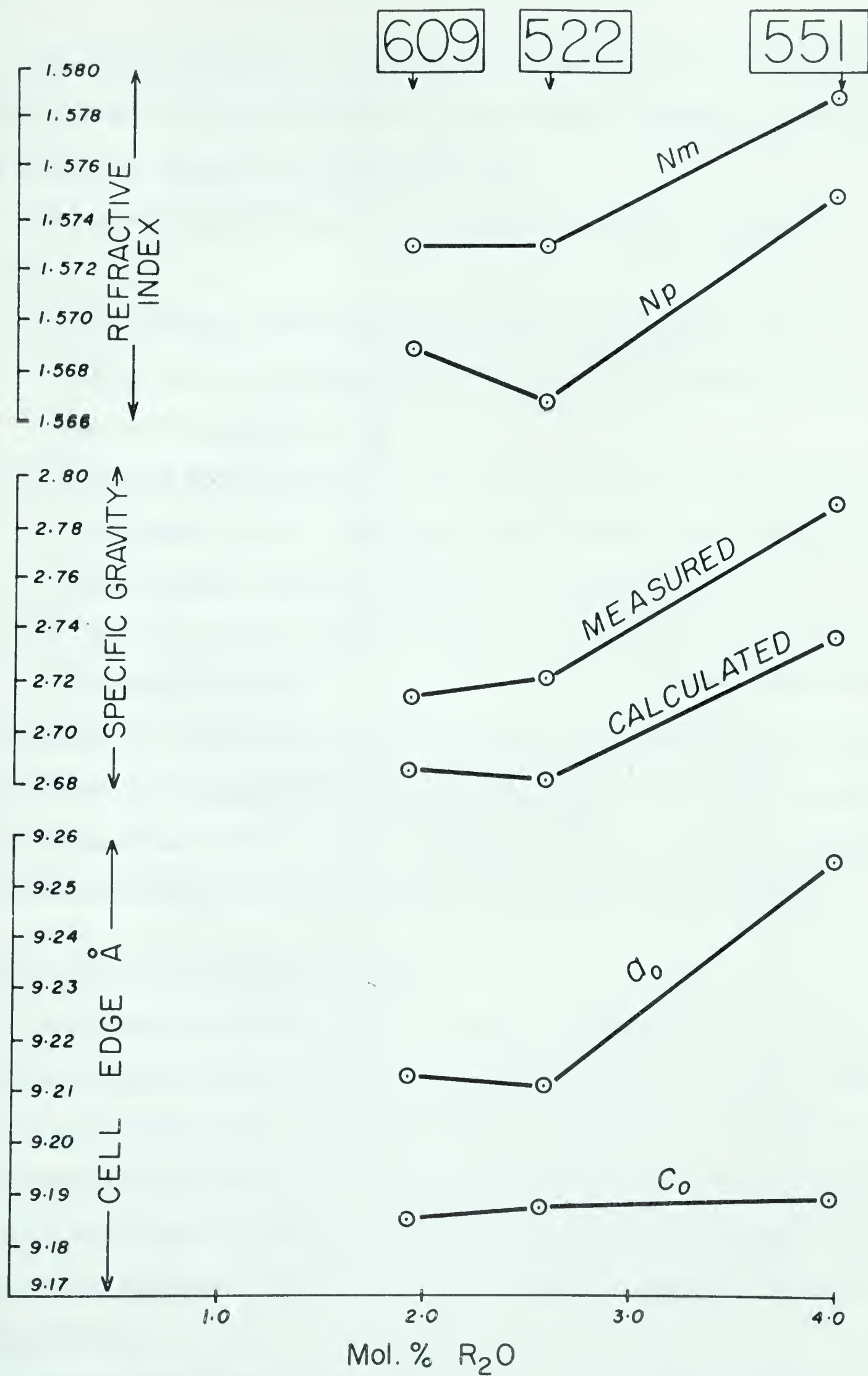


Figure 6. Physical and Chemical Variations of Beryl

All of these substitutions are possible and an estimation of which one has or has not occurred must be based primarily on accurate unit cell parameter measurements and not on chemical composition or structural formulae.

The present study indicates that two and perhaps three distinct classes of beryl exist:

- (1) "C" type beryls where substitution proper occurs and since this is always by cations of larger ionic radii then an expansion of the lattice occurs in the direction of c_0 .
- (2) "A" type beryls where R_2O 's and perhaps higher valence cations reside in the open channels. Once the threshold value of 2.5 mol. % R_2O has been exceeded then the lattice expands in the direction of a_0 .
- (3) "A-C" type beryls probably exist and these involve an expansion in both component directions.

The non-recognition of these fundamental types of beryls is the main reason for the confusion and debate in the geological literature on beryl to date. These different types of beryls can be explained in terms of, firstly, supply and demand of component elements, and secondly, varying physico-chemical conditions at the time of crystallization.

Chemical Variations of the Birch Portage Beryls

Two representative beryl crystals from the Birch Portage area were analyzed chemically, one sample being green and the other yellow. Often the beryls are zoned from green in the crystal centres to yellow towards the crystal margins. This occurs mainly in association with alkali feldspars and the variation of the chemical composition of the two crystals is assumed to represent the change of composition of the last liquid from which beryl and feldspar crystallized. The full chemical composition of the beryls is given in Table 21.

The series of significant chemical changes in the beryl crystals from the

green (551) inner zone toward the yellow (522) outer zone as follows:

TABLE 24

	<u>Green (ppm)</u>	<u>Yellow (ppm)</u>
Li_2O	4	23
Na_2O	3,200	2,400
K_2O	900	500
Rb_2O	98	12
Cs_2O	30	2
H_2O	11,500	7,400
BeO	124,900	100,800

With the exception of lithium, all alkalis show a marked decline with time and this is probably related to the fact that the beryls were co-precipitating with alkali feldspars and there would thus be competition between beryl and feldspar for alkalis. Since alkalis enter the feldspar structure more easily and are essential for its formation, it seems probable that feldspar growth depleted the last liquid of its alkali content.

Cesium however is not very acceptable in feldspars and one might imagine its concentration in the very last liquid, however the BeO content also declined, probably indicating that the liquid was being depleted of these elements by beryl crystallization. It might also be expected that the H_2O content would be higher in the later beryls. However it has already been shown that there is a strict relationship between water and alkalis in the channels of beryl. Thus since the alkali content of the beryls decreases, so does the H_2O . Lithium is the only element which increases but the small ionic radius of lithium makes it unacceptable into feldspars and it seems sensible to find the lithium content rising in the last liquid. It might be expected that lithium, being a small ion would be readily acceptable into the beryllium lattice sites, but the small concentration of lithium in the beryls has no apparent effect and it is not possible to determine whether or not the lithium resides in the channels or in the beryllium lattice positions.

CHAPTER 5

SEQUENCE OF EVENTS

A full appreciation of the paragenesis of beryl pegmatites cannot be obtained without a review of generalized sequence of events in the magmatic geochemical cycle of beryllium. The best, up to date account is given by Beus (1962), and a summary of pertinent factors is given in Appendix II for the convenience of readers.

Origin of the Beryllium-rich Fluid

The origin of the granitic beryliferous fluid is very speculative but the distributions of related pegmatites and their essential similarity, especially those to the southwest near Hanson Lake, indicates a common origin. Simple geometry then implies a deep source and since the rocks are highly differentiated probably represents the concentration of beryllium in a residual phase following strong magmatism during the Churchill Orogenesis.

Since the clark of beryllium in the crust of the earth is 7 ppm compared to 140,000 ppm in beryl, it is most probable that differentiation and magmatism continued for a very long time and was possibly continuous in depth from the time of formation of the Birch Portage country rocks up to the time of beryl crystallization.

Pre-Igneous History

The evidence of Byers (1954) on the Amisk-Missi metasedimentary and meta-volcanic rocks plus the chemical and field evidence of the Birch Portage country rocks suggests deposition of material into a rapidly deepening marine environment so that graywackes were deposited and at least 35,000 feet of sediment, perhaps 50,000 feet.

Termination of sedimentation may have been accompanied by the onset of regional compressive stresses, as it appears that in geosynclines (e.g. Appalachian), sedimentation does not exceed 50,000 feet without some structural readjustments. The compressive stresses led to folding and recrystallization of the sedimentary rocks. The

disposition of the antiform on Birch Portage suggests that the long axis of maximum compression lay in a northwest-southeast direction. Thus the principal stress axes would lie at some angle to this direction, probably less than 60° , and observations in the field suggest that the axes could have been approximately north-south and east-west. These could give rise to shears and in particular, the Sturgeon Weir Fault (north-south) which truncates the antiform. If this interpretation is correct then the probable direction of displacement along the Sturgeon Weir wrench fault is west side south. Evidence does not indicate large displacements.

The east-west principal stress axis may have produced a series of small shears which became filled with magma and following enlargement of the shears, crystallized as dyke-like bodies with long straight sharp contacts.

During compression and folding along a northwest-southeast axis a probable series of extension fractures opened up parallel with the axis of maximum compression, in the more competent country rocks. These became filled with fluids, which eventually crystallized as dyke-like bodies.

Evidence indicates that all these structural adjustments took place in a sequence, the intrusion of dykes after the main folding and development of the Sturgeon Weir fault, as no dykes have been identified east of the fault. The basic dykes were injected first as these show the highest degree of recrystallization and cross-cutting relationships. Felsites were probably intruded after the main recrystallization of the basic dykes as they show lower recrystallization features. Further structural movements formed fracture cleavages in the basic and felsitic dykes, and opened up a younger series of fractures which became occupied by the porphyries, aplites and pegmatites, which are unmetamorphosed.

The intrusion of aplites and pegmatites was into a fan-like fracture system with a dominant southwesterly dip. Structural analysis has shown that the high and low angle dips were dynamically not wholly suitable for pegmatite formation as the aplites are restricted

to high and low dips. While the barren pegmatites may dip at angles varying from 0° to 90° , it is clear that the last beryllium rich fluids favoured intermediate dipping pegmatites, perhaps indicating that the barren pegmatites occupying high and low angle dips are slightly older in age or perhaps cooled more quickly.

Igneous History

The younger dykes are unmetamorphosed and it is possible to derive more information from them compared with the older metamorphosed diabases and felsites.

The aplites and pegmatites were intruded into a fracture system similar to the basic and felsitic dykes, perhaps at a slightly later stage. Evidences of the chemistry and mineralogy of the aplites and pegmatites shows that they were intruded into fractures perhaps held open by the regional stress system, at a depth somewhere near 50,000 feet. Under these conditions crystallization was slow and pegmatites could form. An initial crystallization of aplite sealed up the system and precipitation of minerals began at a temperature near 650°C and terminated near 560°C , under a salt-water vapour pressure of approximately 4,000 bars.

Crystallization was cotectic and subsolvus so that 2 to 3 minerals can be seen in co-existence. There is petrographic evidence to suggest that the centers of the dykes were liquid (parallelism of crystal growth) and that there was a continuous passage of material during crystallization (rhythmic banding). The general unavailability of free cations with high valence (X-ray fluorescence scans show none) prevented the capture of beryllium by the silicate lattices of the feldspars. The presence of about 200 ppm TiO_2 in the pegmatite analyses can probably be explained by titaniferous magnetite which results mainly by recrystallization of country rock inclusions - the country rocks show a TiO_2 content of 5-6000 ppm.

The continued passage of material through the centres of the dykes acted as a cleansing or collecting agent for beryllium until at a late stage a beryllium-alkali-volatile complex probably formed and collected in certain discrete pockets in the dykes,

probably pressure lows at higher elevations (?). Under these conditions the beryllium content was sufficiently high and it began to build its own lattice. However, beryl is always found in association with feldspar and zoned beryls are found intimately associated with feldspar. The change in composition of the last liquid with time is represented by the zoned sequence in the beryl and this reflects an essential partitioning of the remaining elements between beryl formation in the complex probably salt rich phase (beryl is noted for its liquid inclusions of salts) and feldspar in the simpler silicate phase.

After final crystallization of the beryl pegmatites diffusion occurred within the feldspar for a temperature interval of at least 150°C. During this time orthoclase exsolved albite and finally orthoclase inverted almost wholly to microcline. The rate of temperature decrease from 560°C to about 300°C was probably very slow.

Post-Igneous History

The cooling and crystallization of beryl pegmatites marks the last igneous and metamorphic event in the area. Following symmetry inversions within the pegmatite feldspars was probably simple erosion of the overlying rocks and isostatic response whereby the Birch Portage area of Saskatchewan was elevated to its present position.

REFERENCES

- Baadsgaard, H. (1960) "Procedures. Rock and Mineral Analysis." University of Alberta, Rock Analysis Laboratory.
- Bakakin, V.V. and Belov, N.V. (1962) "Crystal Chemistry of Beryl." *Geochemistry* No. 5, pp. 484-500.
- Beus, A.A. (1962) "Beryllium". W.H. Freeman and Company, San Francisco, U.S.A.
- Bowen, N.L. (1937) "Recent High Temperature Research on Silicates and Its Significance in Igneous Geology." *American Journal of Science*, v. 33-34, pp. 1-22.
- Bowen, N.L. and Tuttle, O.F. (1950) "The System $\text{NaAlSi}_3\text{O}_8$ - KAlSi_3O_8 - H_2O ." *Journal of Geology*, v. 58, pp. 489-511.
- Bragg, W.L. and West, J. (1926) "The Structure of Beryl." *Proc. Royal Soc. London*, Series A, v. 3, pp. 691-714.
- Buerger, M.J. and Azaroff, L.V. (1958) "The Powder Method." McGraw Hill Book Company, Inc., New York.
- Byers, A.R. and Dahlstrom, C.D.A. (1954) "Geology and Mineral Deposits of the Amisk-Wild Nest Lakes Area, Saskatchewan." Province Sask., Dept. Mineral Resources, Geology Branch.
- Cameron, E.N., Rowe, R.B. and Weiss, P.L. (1953) "Fluid Inclusions in Beryl and Quartz from Pegmatite in the Middle Town District, Connecticut." *American Mineralogist*, v. 38, no. 3-4.
- Damon, P.E. and Kulp, J.L. (1958) "Excess Helium and Argon in Beryl." *American Mineralogist*, v. 43, pp. 443-459.
- Deer, W.A., Howie, R.A. and Zussman, J. (1962) "Rock Forming Minerals." Longmans, Green and Co., Ltd., London.
- Fleischer, M. and Stevens, R.E. (1962) "Summary of New Data on Rock Samples G-1 and W-1". *Geochemica et Cosmochemica Acta*, v. 26, pp. 525-543.

- Folinsbee, R.E. (1941) "Optical Properties of Cordierite in Relation to Alkalis in the Cordierite-Beryl Structure." *American Mineralogist*, v. 26, pp. 485-500.
- Ginzburg, A.I. (1955) "On the Question of the Composition of Beryl." *Trans. Miner. Mus. Acad. Sci. USSR*, v. 7, p. 56.
- Goldschmidt, V.M. (1954) "Geochemistry." Oxford Claredon Press, England.
- Griffith, W.R., Larrabee, D.N. and Norton, J.J. (1962) "Beryllium in the United States." Map Mr-35, U.S.G.S.
- Heinrich, E.W. (1962) "Geochemical Prospecting for Beryl and Columbite." *Economic Geology*, v. 57, pp. 616-619.
- Hill, W.U. (1934) "Structure of Vitreous BeF_2 ." *Zeits Kryst. (A)* v. 89, no. 481.
- Hillebrand, W.F., Lundell, G.E.F., Bright, H.A. and Hoffman, J.I. (1959) "Applied Inorganic Analysis." John Wiley and Sons, Inc., New York.
- Huttenlocher, H., Husi, Th. and Nowacki, W. (1954) "Roentgenographische und Spektrographische Untersuchungen am Bassit." *Schweiz Mineral. Petrogr. Mitt.*, v. 34, pp. 501-504.
- Lowdon, J.A. and Stockwell, C.H. (1962) "Age Determinations and Geological Studies." Geological Survey Canada, Paper 62-17, Ottawa.
- MacKenzie, W.S. (1954) "The Orthoclase-Microcline Inversion." *Mineralogical Magazine*, v. 30, pp. 354-366.
- Mason, B. (1960) "Principles of Geochemistry." John Wiley and Sons, Inc., New York.
- Norrish, K. (1950) "An X-ray Study of Beryl." *Journal Roy. Soc. West Australia*, v. 34, pp. 1-16.
- Orville, P.M. (1957) "Feldspar Investigations." *Ann. Rept. Director Geophy. Lab.* pp. 206-209.
- Orville, P.M. (1963) "Alkali Ion Exchange Between Vapour and Feldspar Phases." *Amer. Jour. Science*, v. 261, pp. 201-237.
- Pettijohn, F.J. (1948) "Sedimentary Rocks." Harper and Brothers, New York.

- Pyke, M.W. (1962) (in) "Summary Report of Geological Surveys Conducted in the Precambrian Area of Saskatchewan." Dept. Min. Resources, Regina, Sask.
- Pyke, M.W. (1963) as above.
- Rankama, K. and Sahama, Th.G. (1950) "Geochemistry." University of Chicago Press.
- Reynolds, R.C. (1963) "Matrix Corrections in Trace Elements Analysis by X-ray Fluorescence: Estimation of the Mass Absorption Coefficient by Compton Scattering." *American Mineralogist*, v. 48, pp. 1133-1143.
- Schaller, W.T., Stevens, R.E. and Jahns, R.H. (1962) "An Unusual Beryl from Arizona." *American Mineralogist*, v. 47, pp. 672-699.
- Smith, J.V. (1960) "Phase Diagrams for Alkali Feldspars." Int. Geol. Congress, XXI Session, Part XXI, Copenhagen.
- Sosedko, T.A. (1957) "The Change of Structure and Properties of Beryls with Increasing Amounts of Alkalis." *Mem. All Un. Min. Soc.*, v. 86, pp. 495-499.
- Spiegel, M.R. (1961) "Statistics." Schaum Publishing Co., New York.
- Thornton, C.P. and Tuttle, O.F. (1960) "Chemistry of Igneous Rocks. I. Differentiation Index." *Amer. Jour. Science*, v. 258, pp. 664-684.
- Turner, F.J. and Verhoogen, J. (1960) "Igneous and Metamorphic Petrology." McGraw Hill Book, Inc., New York.
- Tuttle, O.F. and Bowen, N.C. (1958) "Origin of Granite in the Light of Experimental Studies." *Memoir 74, Geol. Soc. America.*
- Van Valkenburg, A. and Weir, E. (1957) "Beryl Studies $3\text{BeO} \cdot \text{Al}_2\text{O}_3 \cdot 6\text{SiO}_2$." *Bull. Geol. Soc. America*, v. 68, pp. 1808.
- Vance, J.A. (1961) "Polysynthetic Twinning in Plagioclase." *American Mineralogist* v. 46, pp. 1097-1119.
- Vokes, F.M. (1958) "Beryllium in Canada." Map 1045A-M2, G.S.C. Publication, Ottawa.

APPENDIX I

PHYSICAL AND CHEMICAL TECHNIQUES

SAMPLE PREPARATION

Thirty-one samples of rocks and minerals were disintegrated in a Brown Jaw Crusher. Each sample was coned and quartered and a representative coarse and fine fraction from each of the four quarters was placed in a separate jar and used as the stock sample. All of the latter samples were ground for 2 minutes in a Pica Blender Ball Mill and the sample extracted from the centre of the cylinder was used for physical and chemical measurements as it is least affected by iron contamination. A second series of samples were taken from the original stock and ground by hand in an agate mortar. These were used in X-ray fluorescence analysis along with the Pica Blended samples to test contamination and reproducibility.

The three analyzed beryl crystals were prepared differently. They were all large single crystals and after a preliminary disintegration in the jaw crusher, were picked by hand to ensure a separate completely free from feldspar as the alkali elements are an important consideration in the crystal chemistry of beryl. Two of the beryl crystals are from Birch Portage, one being yellow and the other green and represents the zonation of beryl from green to yellow that is usually seen in association with feldspar. The third sample was a comparatively pure aquamarine from Mt. Antero, Colorado, U.S.A.

The clean fragments of beryl were ground in an agate mortar and fractions used for all physical and chemical measurements.

PHYSICAL MEASUREMENTS

In this section the X-ray diffraction and fluorescence analyses, specific gravity and refractive index determinations are described.

X-ray Diffraction

The finely ground samples of beryl and feldspar were smeared on a glass slide by means of a duco-acetone mixture and scanned at a speed of $1/4^\circ$ minute on a Norelco Diffractometer Type 12045B/3 using a Geiger Tube pick-up, a Cu K_{α} radiation source energized by 35 KV and 15 MA, and 1° slits. Three samples of each beryl were scanned three times and the peak positions calibrated by the silicon powder standard briquette. At the time of measuring the instrument was well aligned and observed 2θ angles corresponded with recommended d spacings to within $\pm 0.01^\circ 2\theta$ for silicon metal.

X-ray diffraction runs were made to obtain the following information on beryl which is fully discussed in Chapters 3 and 4.

- (1) The unit cell dimensions of the beryl crystals.
- (2) Completely indexed powder diffraction data of beryl which is as yet, unsatisfactorily expressed in the geological literature.
- (3) Alkali feldspars were also studied by diffraction techniques to determine the degree of triclinicity and the content of $\text{NaAlSi}_3\text{O}_8$ remaining in solid solution with KAlSi_3O_8 .

X-ray Fluorescence

The Pica Blended samples were diluted with cellulose powder using a volume to volume relationship. This is approximately 9:1 sample : cellulose, by weight. The mixture was homogenized by agitating for 20 minutes on the Pica-blender. The samples ground in the agate mortar were not diluted.

All samples were briquetted in an Applied Research Laboratories Inc. Briquetting machine Type 4451, for 60 seconds at 30,000 p.s.i. The samples were backed and rimmed with cellulose powder for strength and the shape of the beryl powder cross-section was made rectangular rather than cone shaped. The reasons for this are:

- (i) A uniform thickness and volume of sample are presented to the X-ray machine in every case.

- (ii) Since disintegration of briquettes occurs in most cases within the sample powder and not between the sample and cellulose, coned powder is not put in a state of lowest free energy by the applied differential stress when the maximum principal stress axis is parallel with the smaller dimension of the briquette. Strain occurs mainly by elastic strain rebound following stress release and this is higher for hard and monomineralic materials.

The powder is put in the state of lowest free energy if three procedures are closely followed:

Firstly, the sample should be rectangular in cross section, so as to minimize differential stress within the gross geometry of the system. Secondly, the press should be scrupulously clean so that the sample-press interface is a minimum friction surface. Thirdly, the pressure should be applied very slowly, especially in the low compression ranges so that the grains can adjust into an interlocking aggregate of least free energy, which is of course, a state of equilibrium. Slow pressure release is also recommended although this is not too critical.

The briquettes were mounted in Norelco X-ray fluorescence equipment, the basic unit, type 12215/0 having a constant potential output. Radiation sources were Mo Target Tube energized by 50 K.V. and 40 M.A.

A.D.P., E.D.D.T. and LiF, analyzing crystals were used in combination with a Scintillation counter operating at 800 volts, a Flow Proportional Counter operating at between 1460-1510 volts and receiving 0.25 S.C.F./H supply of a 90% Methane, 10% Argon gas mixture, and an X-ray path vacuum of 0.2 mm Hg. Pulse Height Analysis was used for the light elements, and counts /100 seconds were used throughout.

Complete recording conditions for individual elements are given in Table 25. The cellulose diluted briquettes gave best values for SiO_2 , Al_2O_3 , TiO_2 and MgO , while the undiluted gave best results for Fe_2O_3 , MnO , CaO , K_2O , SrO , and Rb_2O .

The conversion of counts per second into weight per cent was made with a conventional calibration curve (which depends upon the accuracy of the chemically

TABLE 25

Recording Conditions for X-ray Fluorescence Analysis

<u>Element</u>	<u>Peak</u>	<u>B.G.</u>	<u>XTAL</u>	<u>CTR</u>	<u>CTRV</u>	<u>X-RP</u>	<u>PHA</u>	<u>PHLV</u>	<u>PHW</u>
Mg	107.17	106.50	ADP	F.P.	1500	Vac.	Diff.	4	7
Al	112.55	111.00	EDDT	F.P.	1500	Vac.	Diff.	4	8
Si	77.90	77.20	EDDT	F.P.	1480	Vac.	Diff.	6	8
K	20.10	19.00	EDDT	F.P.	1480	Vac.	Diff.	8	15
Ca	14.65	14.00	EDDT	F.P.	1510	Vac.	Diff.	8	15
Ti	86.08	85.00	LiF	F.P.	1485	Air	Diff.	10	15
Mn	62.80	61.50	LiF	F.P.	1470	Air	Diff.	5	20
Fe	57.35	56.00	LiF	F.P.	1500	Air	Int.	1.5	--
Rb	26.42	25.70	LiF	Sc.	820	Air	Int.	1.5	--
Sr	24.94	25.70	LiF	Sc.	820	Air	Int.	1.5	--

Abbreviations:

B.G.	- Background in $^{\circ}2\theta$ (peak also)
XTAL	- Analyzing Crystal
CTR	- Counter
CTRV	- Counter Voltage
X-RP	- X-Ray Path
PHA	- Pulse Height Analyzer
PHLV	- Pulse Height Level Voltage
PHWV	- Pulse Height Width Voltage
ADP	- Ammonium Dihydrogen Phosphate
EDDT	- Ethylene-Diamine-Ditetra-Acetate
LiF	- Lithium Fluoride
F.P.	- Flow Proportional Counter (gas flow)
Sc.	- Scintillation Counter
Vac.	- Vacuum
Diff.	- Differential (PHW Engaged)
Int.	- Integral (PHW disengaged)

analyzed standards) for all elements with the exception of Rb and Sr. The conversion in this case was made using the Compton Peak correction (Reynolds, 1963). This correction allows for the variations in the mass absorption coefficients of the samples which is a particularly critical factor in the quantitative determination of elements in trace amounts.

A close estimation of the mass absorption coefficient (μ) is made by measuring the intensity of radiation of a Compton Peak which is produced by scattering of the X-ray by the sample with a resultant change in wave length (incoherent scattering). The change in wave length is expressed by the function $\lambda + \Delta\lambda$ and thus occurs at a slightly larger 2θ than the peaks for the X-ray target material, which was in this case molybdenum. Thus the MoK peak occurs $20^\circ 2\theta$ for the LiF analyzing crystal and the Compton Peak at $21^\circ 2\theta$ approximately.

The values of μ were accepted at a wave length of 0.9°A which is close to that of Rb and Sr $K_{\alpha 1}$ radiation, e.g. Sr $K_{\alpha 1} = 0.875\text{A}$, Rb $K_{\alpha 1} = 0.926\text{A}$. G-1 was used as a comparison standard and the following values of Rb and Sr cited in Fleischer and Stevens (1962) were accepted: Rb 220 ppm., Sr 280 ppm. The equation for computation of trace elements in the unknown is:

$$\text{ppm } Z = \frac{\text{cps. } Z \text{ } K_{\alpha 1}}{\text{cps. } G-1 \text{ } K_{\alpha 1}} \times \frac{\mu_{0.9^\circ\text{A}} Z}{\mu_{0.9^\circ\text{A}} G-1} \times \text{ppm } G-1$$

where Z = unknown, μ = mass absorption coefficient, and CPS = counts per second on peak minus background intensities. For rapid computation of a large number of samples the equation can be simplified into:

$$\text{ppm } Z = \text{cps. } Z \text{ } K_{\alpha 1} \times 0.9^\circ\text{A } Z \times \text{constant}$$

where the constant is:

$$\frac{\text{ppm. } G-1}{\text{cps. } G-1 \text{ } K_{\alpha 1} \times \mu_{0.9^\circ\text{A}} G-1}$$

Diluted and undiluted samples were used for all elements and a comparison of the results yields three noteworthy observations:

- (1) The diluted and undiluted yield similar results in general, although the precision of measuring and detectability of the undiluted samples is better than for the diluted.
- (2) The diluted samples give higher Sr and Rb values than the undiluted. The reason for this is not understood, but may be partly explained by the fact that the process of dilution with cellulose differentially alters the relative mass absorptions between unknown and standard.
- (3) Inspection of the SiO_2 and Al_2O_3 analysis of beryl and feldspar indicate that the diluted samples yield more sensible results than those of the undiluted samples, as these latter are higher than the stoichiometric requirements of the formulae. Since the radiation of these two elements is weak, it is consequently adversely affected by absorption by other elements to a varying degree. The process of dilution while decreasing the intensities slightly, tends to produce a similar matrix for standards and unknowns alike and thus decreases the effects of differential absorption.

The precision error of each element was determined from a series of measurements made on the standard G-1 following every three unknowns. The variations observed were statistically analyzed, employing the standard deviation function. According to Spiegel (1961) ± 1 S.D. from the mean includes 68.27% of the number of points N for normal distributions. This value has been accepted as the precision error as it gives 2:1 chance of reliability. Since the value of the S.D. in absolute amounts varies proportionately as the intensity of radiation, then the precision is expressed as a percentage, i.e. $\frac{\text{S.D.}}{\text{Mean}} \times 100$

The estimation of detectability is made on the variation of background of the standard sample and the value computed through the standard deviation function. According to Spiegel (1961) ± 3 S.D. from the mean includes 99.17% of all the points. Thus 3 S.D. above the mean background expressed in WT% is accepted as the detectability (Table 26).

TABLE 26

Statistical Data for X-Ray Fluorescent Analysis

<u>Oxide</u>	<u>Precision (%)</u>	<u>Detectability (wt.%)</u>
MgO	2.25	0.90
Al ₂ O ₃	2.68	0.30
SiO ₂	1.21	0.33
K ₂ O	1.01	0.003 ?
CaO	0.59	0.032
TiO ₂	1.89	0.004
MnO	3.32	0.007
Fe ₂ O ₃	1.19	0.02
Rb ₂ O	0.90	0.0003 ?
SrO	1.40	0.0005 ?

Specific Gravity Measurements on Beryl

Specific gravity measurements were made with a precision quartz glass pycnometer which had been carefully calibrated to variations of 0.1°C and the accuracy is better than ± 0.001 . The principle of operation is that of Archimedes whereby the weight of water displaced is equal to the apparent loss in weight. The author found that while theoretically the specific gravity of small amounts of material are measurable, weights of at least 1 gm should be used in order to give a confident result. In detail, the procedure is as follows:

- (1) Using finely ground powder, dry sample for 8 hours at 110° to remove adherent water. Warm pycnometer and stopper. Allow all to cool.
- (2) Weigh pycnometer and stopper, with stopper out.
- (3) Add 1 gm of sample by means of a small funnel.
- (4) Fill pycnometer half full with previously boiled water and boil the mixture carefully for 10 minutes to remove any air bubbles.
- (5) Allow sample to settle and fill up pycnometer to brim with distilled water. Insert stopper and immerse the open end in a beaker of distilled water.
- (6) Allow to stand for at least 8 hours, immediately adjacent to the weighing scales, to stabilize.
- (7) After 8 hours place pycnometer onto scales, dry wet end of stopper with filter paper and as the water evaporates from the capillary, allow scales to swing free. Accept the weight as the water passes the first graduation mark in the capillary tube of the stopper.
- (8) Remove part of the water from the pycnometer and insert a thermometer clearly graduated to 0.1°C .

An example of the calculation is given below (D.R. 609):

Wt. Pyc. + Stopper + Sample	=	17.8837
Wt. Pyc. + Stopper	=	<u>16.5521</u>
Wt. Sample	=	1.3316 gms
Wt. Pyc. + Stopper + Sample	=	17.8837
Wt. Water in Pyc. at 22.5°C	=	<u>9.2011</u>
Theoretical Wt. Pyc., Stopper, Sample and Water	=	27.0848
Measured Weight	=	<u>26.5957</u>
Wt. Water Displaced	=	<u>0.4891</u> gms
$S.G. = \frac{\text{Mass}}{\text{Volume}} = \frac{1.3316}{0.4891}$	=	2.7226
Correction at 22.5°C	=	-0.0079
True S.G.	= 2.7147	= <u>2.715</u>

The specific gravity was also determined in another manner when the mass was calculated from the experimentally determined chemical composition and the volume from the measurements of the unit cell parameters. In general these results are lower than the measured specific gravity. The results of the specific gravity determinations are given below:

TABLE 27

	<u>609</u>	<u>522</u>	<u>551</u>
S.G. measured	2.715	2.722	2.791
S.G. calculated	2.686	2.681	2.739

Refractive Index Measurements on Beryl

In general the author had considerable difficulty in obtaining optic interference figures mainly as a result of grain size. The following method was thus employed in order to determine orientation:

A completely isotropic section is one which shows no birefringence and consequently can only contain 1 refractive index (N_m). These sections were sought out and N_m measured in monochromatic sodium light using the Becke line technique.

In contrast to the above, thin grains showing strong birefringence must contain both refractive indices. The grains have no habit and the orientation of the vibration directions was made with a gypsum plate in white light, when the stage is rotated until enhancing or retardation gypsum colours appear. The gypsum plate and N_p are both length fast and this direction appears at maximum blue. The grain was then rotated toward the north-south lower vibration direction of the Leitz microscopes to extinction when N_p was measured in monochromatic sodium light.

When the two indices had been bracketed, the immersion oils were mixed in various quantities until the exact R.I. was matched. The refractive index of the oil were measured in absolute terms on the Abbe Refractometer. The results of this investigation, which are given below, are thought to be no worse than ± 0.002 .

TABLE 28

	<u>609</u>	<u>522</u>	<u>511</u>
$N_m (N_z)$	1.573	1.573	1.579
$N_p (N_x)$	1.569	1.567	1.575

CHEMICAL MEASUREMENTS

This study can be divided into two main sections, the spectro-chemical determination of alkalis and the gravimetric determination of beryllium in beryl. The water content of some of the samples was determined by loss on ignition.

Determination of Alkalies

The digestion of samples and production of a liquid containing only Mg, Li, Na, K, Rb, Cs, SO_4 , is essentially that described by Baadsgaard (1960) in the University of Alberta, Silicate Rock Analysis Lab Book. In essence this consists of digestion with HF and H_2SO_4 , evaporating to dryness and leaching of alkalies with hot water. Li, Na, K, Rb, and Cs responses were measured on flame photometers and converted to ppm by the standard type calibration curve. The Perkin-Elmer Flame Photometer has a poor optical resolution and there is a large interference of Na and K when Li, Rb, and Cs are measured. For this reason, the accepted values were those measured on the Beckman D.U. Flame Photometer (Courtesy Soil Science Department, U. of A.) which has an excellent optical resolution and a reported precision detectability of 1 ppm. The determined recording conditions using an oxy-acetylene flame are given below in Table 29.

TABLE 29

	<u>Li</u>	<u>Na</u>	<u>K</u>	<u>Rb</u>	<u>Cs</u>
Phototube	red	blue	red	red	red
Wave Length	670.8	588.5	757	630	852.1
Resistor	3	2	3	3	3
Sensitivity	0	6	0	0	0
Zero Supp.	0	0	0	0	0
Slit Width (mm)	0.8	0.03	0.3	1.2	1.0
Acetylene pressure (lbs)	6	3	2.5	7	6
Oxygen pressure (lbs)	12	12	9	10	10

There was good agreement between the results of Na and K for Perkin-Elmer and Beckman D.U. Instruments, although the Perkin-Elmer values were a little higher probably due to mutual interference of sodium and potassium. The K and Rb results

determined by X-rays were similar to those of the Beckman D.U. Photometer. Li, Na, and Cs values obtained on the Beckman instrument were accepted as correct.

Determination of Beryllium in Beryl

The quantity of BeO in the three beryls was determined by gravimetric means. In general the analytical procedure consists of removing all the other elements from solution, destroying organic matter and then precipitating Be and $\text{Be}(\text{OH})_2$ and igniting to BeO at 1000°C on a Meker burner. The problem is much complicated by the similar behaviour of Be^{2+} and Al^{3+} in solution and one must carefully control the chemical conditions in order to separate them fully. It is a fairly well recognized fact that the spectrochemical analysis of Be in trace amounts is exceedingly difficult. However in the opinion of the writer an accurate gravimetric analysis of BeO in macroquantities is equally difficult due particularly to the coherent behaviour of Be and Al.

The analytical procedure is based on Baadsgaard (1964, Personal Communication) and on Hillebrand's Applied Inorganic Analysis (1957) with modification where necessary. The general schematical procedure outlined below must be followed in detail as Deer (1962) shows up to 3% impurities of all types characterizing beryl.

- (1) Na_2CO_3 fusion of sample HCl. Remove silica by double evaporation.
- (2) ppt. BeO and R_2O_3 with NH_3OH when Mg, Ca etc. and alkalis remain in solution.
- (3) Dissolve precipitate in HCl and precipitate Fe, Ti, Zr, etc. with NaOH.
- (4) Precipitate Al from filtrate with 8-hydroxyquinoline.
- (5) Evaporate filtrate to dryness and destroy organic matter with conc. HNO_3 .
- (6) Precipitate Be as $\text{Be}(\text{OH})_2$ with NH_3OH and ignite until $\text{Be}(\text{OH})_2$ reduced to BeO. Weigh.

Removal of Si

Approximately 0.8 gms of finely ground beryl powder were placed in a platinum crucible together with about 4 gms of Na_2CO_3 . The contents were thoroughly mixed

with a glass rod and then a little extra flux was added as a thin layer to reduce "spattering". The crucible was then covered and heated up very slowly in small stages over a Tirrel burner. After 15 minutes of slow heating up, the crucible was given full heat for 10 minutes. No spattering onto the crucible cover occurred. The crucible was heated for 30 minutes over a Meker burner to complete fusion and as the molten mass began to cool it was spread over the walls of the crucible by gentle rotation of the crucible. After cooling the crucible was rapidly heated up to dull red heat and then cooled on the marble slab. This latter process greatly reduces the cohesion of the (Sample- Na_2CO_3) cake with the platinum crucible.

Distilled water is then added to the crucible and after 5 minutes the cake was easily and completely removed into a 250 cc porcelain casserole. The casserole was covered with a ribbed watch glass 10 cc. of conc. HCl slowly added by means of curved funnel. 2 cc. conc. HCl were added to the crucible to remove any last traces of carbonate and the solution poured into the casserole via the funnel. Crucible and funnel were washed with 5 cc. distilled H_2O . By these means chlorides of all the metals are formed by decomposition of carbonates and silicic acid is formed. Since no precipitate could be seen after the reactions had ceased then it is obvious that all SiO_2 was in solution.

The solution was evaporated to dryness on a water bath when all of the chlorides and silicic acid crystallized. The resultant solids were then carefully broken up into very small pieces by means of a glass rod and 5 cc. conc. HCl added to dampen all solids. This prevented the formation of insoluble basic salts of iron when 40 cc. H_2O were added. 15 minutes heating dissolved all the metal salts. Since the silicic acid is a very fine precipitate, the solution was allowed to stand overnight prior to filtering so that most of the filtrate was decanted rather than filtered through a 9 cm. Black Band (Schleicher and Schull) filter paper. The filtrate, caught in 250 cc beaker, still contains some silicic acid and the process of evaporation was repeated, using the

same casserole, and filtered through the same filter paper.

Removal of Alkalies, Ca and Mg

Four gms of NH_4Cl were added to the filtrate in the 250 cc beaker and the solution diluted to 200 cc. Most of the free HCl was neutralized with NH_3OH . The solution was then heated to boiling and ammonia added dropwise with continuous stirring. Using methyl red as the indicator, ammonia was added until the solution just turns yellow at pH 6.5. At this point, total precipitation of the R_2O_3 groups (+Be) was complete. The solution was boiled for 1 minute and then allowed to cool.

A 11 cm. white band filter paper was used and the first filtrate examined for purity. The beaker and precipitate were washed several times with hot 2% NH_4NO_3 . The precipitate was retained.

Removal of Fe, Ti, Mn, Zr, P, etc.

The funnel, filter paper and filtrate from the above, after several washings with hot 2% NH_2NO_3 were digested with 20 cc hot 1:1 H_2SO_4 and the solution diluted to 100 cc. The solution was then almost neutralized with sodium hydroxide and heated to boiling point. This was then poured slowly with constant stirring into 100 cc of a hot, fresh 10% solution of NaOH , boiled for 5 minutes and allowed to settle out. Prior to filtering the 11 cm white filter paper was washed with a hot 5% NaOH solution containing a little Na_2SO_4 . This latter solution, which sealed the pores in the filter paper was also used to wash the precipitate. It is recommended not to use ammonium salts in this operational stage.

Removal of Al

The separation of Al and Be is perhaps the most difficult stage of the operation and special care must be taken to ensure that no Be precipitates with Al and also that all Al is precipitated otherwise it will come down with the Be. For complete separation pH 6 is recommended and experience showed that if the solution was more basic than this, then Be became complexed with 8-hydroxyquinoline (see Discussion of Results).

The filtrate from the previous operation was made slightly acid with 1:1 HCl and diluted to 300 cc as it is important that the conc. of Al does not exceed 0.1 gms per 100 cc. The acetic acid solution of the quinolate reagent was prepared by dissolving 10.5 gms of 8-hydroxyquinoline powder in 24 cc of glacial acetic acid. Warming almost completely dissolved the powder. After cooling the solution was filtered through an 8 cm white filter paper, diluted to 200 cc, and divided into 3 portions of 66 cc, one for each sample. 1 cc of this solution will precipitate 0.0058 gms of Al_2O_3 (Hillebrand, 1959).

The beryllium filtrate was warmed to 60°C and a 59% excess (66 cc) of 2N solution of ammonium acetate was added until a permanent precipitate of aluminium quinolate occurred. 20 cc more was added to ensure complete precipitation. After settling the mixture was filtered through a glass frit into a vacuum flask and the precipitate dried at 110°C. The dried precipitate was weighted as $\text{Al}(\text{C}_9\text{H}_6\text{ON})_3$ containing 11.10% Al_2O_3 . This weight % of Al_2O_3 was compared with that determined by X-ray fluorescence.

Precipitation of Be

The filtrate from the above was treated with 5 gms of NH_4Cl as a buffering agent and the solutions evaporated to complete dryness on a steam bath. 10 cc HNO_3 were added and an immediate reaction occurred with the evolution of NO_2 . After 6 hours of continuous heating 10 cc conc. HNO_3 were again added and the process repeated until no further reaction occurred. At this point, all the organic matter has been destroyed and only (H^+) $(\text{NO}_3)^+$ and $(\text{Be})^+$ remained.

The residual solution was diluted to 200 cc and heated on a steam bath until all the fine solids completely dissolved. The volume was then increased to about 600 cc and the solution raised to boiling point. Methyl red indicator was added and NH_4OH dropwise until the solution just turned yellow, on continuous stirring. At this point $\text{Be}(\text{OH})_2$ began to precipitate. It is important to have an excess of NH_4OH to ensure complete precipitation of Be and $\text{Be}(\text{OH})_2$. However if the solution becomes more

alkaline than pH 10 then the $\text{Be}(\text{OH})_2$ becomes partly soluble again. Thus as soon as the methyl red indicator turned yellow at pH 6.5, a quantity of phenolphthalein was added. This remained colourless on continuous stirring and adding of NH_4OH until at pH 9.5 it turned red. At this point the precipitation of $\text{Be}(\text{OH})_2$ was complete and an excess of NH_4OH had been added without any $\text{Be}(\text{OH})_2$ passing back into solution.

The precipitate was filtered off after cooling and washed very carefully with 2% NH_4NO_3 solution, rather than pure distilled water as $\text{Be}(\text{OH})_2$ is partly soluble in almost pure water.

The filter paper and precipitate were loosely folded up, placed in pre-weighed covered crucibles, and slowly warmed up over Tirrel burners with cover slightly displaced. After the water had evaporated the temperature was very slowly increased in stages until all the filter paper had burned off. At this point the crucibles were transferred to Meker burners and ignited at a 1000°C for one hour to reduce $\text{Be}(\text{OH})_2$ to BeO . After cooling in a dessicator the weights were recorded. The ignition was repeated for 30 minutes and cooling and weighing showed a difference of only 0.0001 gms. These second values were accepted as correct.

Discussion of Results

The measured BeO content for the three beryls are as follows:

TABLE 30

	<u>BeO</u>	<u>Analyses Total</u>
609 Aquamarine	7.75%	96.22
522 Yellow Beryl	12.49%	99.36
551 Green Beryl	10.08%	99.11

The maximum stoichiometric requirement of BeO in the standard formula is 13.96%. Deer(1962) shows a variation of BeO from 10-14% and since 522 and 551 are within this range, and their analyses totals near a 100% then these results are good.

Comparison of the Al_2O_3 measured as quinolate with X-ray fluorescence are within the experimental limits for these latter two.

The analyses of 609, the comparison standard is obviously incorrect. It was noted during procedure (Removal of Al) that the precipitate was heavier and darker than for the other two. At that time it was thought that the solution might be too basic and some of Be having precipitated with Al had become complexed with quinolate. Weighing of this Al precipitate as aluminium quinolate show that it was heavier than the Al_2O_3 total determined by X-rays and outside the normal limits of Al_2O_3 for beryl. Accepting the Al_2O_3 content measured by X-rays as correct, calculation showed that the quinolate was heavier by an amount equivalent to 3% BeO. It is not possible to more specific than 3% as the X-ray determination of Al involves a precision error of 2.68%. In the concentration range of 20% this is equivalent to $\pm 0.5\% \text{Al}_2\text{O}_3$.

Thus 3% was accepted as the amount of BeO lost during the removal of Al and the total of BeO has been increased from 7.75% to 10.75%. This raises the total of the analyses to 99.22% and the BeO content to within the experimentall determined range for beryl. It might be noted that this particular sample (609) has no bearing whatsoever of the writer's interpretation of the paragenesis of the Birch Portage Beryl Pegmatite Deposit.

APPENDIX II

GEOCHEMISTRY OF BERYLLIUM

Abundance and Distribution

Beryllium is decidedly a lithophile element and comparison with meteorites leads to the conclusion that it is concentrated in the earth's crust. Its average abundance in the crust is 7 ppm (Rankama, 1950) or 0.2 atoms of Be/10,000 atoms of Si (Mason, 1960). The abundance of beryllium, like boron and lithium is very low considering its low atomic number as in general the lower the atomic number of an element, the greater its abundance. Goldschmidt (1954) believes that the reason for this to lie in the origin of the earth when prior to its birth as a planet, the earth passed through a stage of high stellar temperatures normally associated with stars and suns. Under these conditions the beryllium nucleus becomes unstable against collision with thermally accelerated protons and breaks down.

Beryllium is thus a very rare element indeed and the geochemical stages and conditions through which beryllium must pass, requires a concentration of 20,000 times to produce a sufficient concentration to build its beryl lattice with silicon, aluminium and oxygen.

General Properties of Beryllium

Beryllium belongs to Group II of Mendeleev's periodic system and has an atomic radius of $1.13\text{\AA}$. The ionic radius of the bivalent ion of beryllium is $0.34\text{\AA}$. The ion is highly symmetrical, has the structure of the noble gas type, and is characterized by the lowest polarizability among ions of this type and by a considerable (second after silicon) polarizing power.

The small radius of beryllium ion Be^{2+} (smallest among metals) strengthens the bond between its valence electrons and the nucleus and is one of the causes of the great stability of the lattice of beryllium minerals. For the same reason beryllium has a large

electronegativity, which approaches that of aluminium. Because of its chemical characteristics (e.g. degree of dissociation of its compounds, etc.) beryllium, like aluminum, occupies an intermediate position between the typical cations and the complex forming elements. Covalent bonds play an important role in its oxygen compounds. The amphoteric properties of beryllium are manifested especially strongly in its oxides and halides of the alkali metals, in which beryllium plays the role of a complex-forming element and forms various beryllates.

The co-ordination member of beryllium is 4, and in its natural compounds it is always surrounded by four oxygen ions forming compact $(\text{BeO}_4)^{6-}$ tetrahedra.

A remarkable feature of the beryllium-oxygen tetrahedron is its resemblance to silicon-oxygen tetrahedron (SiO_4) which has nearly the same parameters. There is also a structural resemblance between the beryllium-oxygen and the aluminium-oxygen tetrahedra, although in this case the difference is somewhat greater. The similarity of the $(\text{BeO}_4)^{6-}$, $(\text{AlO}_4)^{5-}$, and $(\text{SiO}_4)^{4-}$ tetrahedra plays an exceptionally important role in the geochemistry of beryllium, determining the character of its isomorphism and its double role in silicate structures.

The characteristic of beryllium during the magmatic stage of formation of acid and alkalic rocks are explained by the manner of its diadochic entry into the crystal structures of various silicates, the main factor being the resemblance of the $(\text{BeO}_4)^{6-}$, $(\text{SiO}_4)^{4-}$ and $(\text{AlO}_4)^{5-}$ tetrahedral groups. The similar behaviour of Be and Si is exemplified by considering the structural resemblance between polymorphs of BeF_2 and different modification of SiO_2 (Hill, 1934). At present three polymorphs of BeF_2 are known, two of which are structural analogous to cristobalite and the third, obtained by crystallization of BeF_2 in the presence of NaF between 425° and 528°C, exhibits a clear structural resemblance to low quartz.

The most important factor preventing the capture of beryllium by the silicates is the difference in the electrostatic charges of Be and Si, which requires that a cation of high charge enter the lattice at the time of substitution of $(\text{BeO}_4)^{6-}$ for $(\text{SiO}_2)^{4-}$ in

order to preserve its electrostatic equilibrium. Thus, the possibility of substitution of $(\text{BeO}_4)^{6-}$ for $(\text{SiO}_4)^{4-}$ depends to a large extent on the presence or absence during the process of crystallization of free cations with high valence which could restore the equilibrium of the silicate lattice disturbed by the substitution of Be^{2+} for Si^{4+} and the compensate for the energy loss involved in such a substitution. The nature of substitution is therefore heterovalent.

Beryllium in the Magmatic and Pegmatitic Process

In considering the normal course of crystallization of granitic melts, it is necessary to note a very important fact which affects the history of beryllium dispersed through the granitic magma. During the early stages of crystallization of granite, the titanium and rare earths contained in the melt may become bound in ilmenite and monazite. The negligible beryllium content in the granitic melt precludes the possibility of formation of beryllium minerals during the magmatic stage and the unavailability during the main stages of crystallization of sufficient amounts of free cations with high valence inhibits the capture of beryllium by the lattice of the essential minerals of granite.

The limited dispersion of beryllium in the early minerals crystallizing from granitic magma results in its concentration in products of the final stages of crystallization in pegmatites, hydrothermal and pneumatolytic rocks. The increased beryllium content in late minerals of granitic rocks is characteristic and muscovites can contain up to 50 ppm Be. It is genetically significant to note the common association of extensive beryl pegmatites fields with intrusion of biotite granites characterized by a limited development of the late muscovite bearing facies.

During the pegmatitic process, the granitic magma may become enriched in beryllium if the factors favouring its dispersion in granite or passage into the pneumatolytic phase are absent. At a late stage, the concentration of beryllium may become high enough that beryllium builds its own lattice with silicon and aluminum. Generally

in rocks of these types the earlier paragenetic pegmatite zones e.g. graphic to perthitic, contain a lower concentration of beryllium than the parent granite. For example, Beus (1962) shows that the Be content of graphic granite averages 3 ppm and microcline perthite 2 ppm compared to 5 ppm in granites of the U.S.S.R. Spectrochemical analyses of 4 Birch Portage central phase pegmatite samples containing visible beryl show comparatively enormous concentration i.e. 40, 50, 300, and 500 ppm BeO (Pyke, 1962).

It is possible that the late pegmatitic crystallization of beryl occurs under conditions of strong over-saturation with silica and the accumulation of volatiles (mainly H_2O), and continues throughout the pneumatolytic-hydrothermal stage of replacement. During these stages the characteristics of the migration of beryllium are probably determined by two geochemical factors:

- (1) The continuously increasing concentration of the volatiles (OH, F, Cl, CO_2) in a virtually closed system.
- (2) The high concentrations of alkalis (mainly sodium).

It is possible that under these conditions the highly mobile complex compounds such as chlorine, fluorine, and carbonate beryllates of the alkalis form. As the pegmatite crystallizes, these migrate in gaseous form, or as aqueous solutions, into the still uncrystallized central and upper parts of the pegmatites. That these intermediate compounds can exist is strongly indicated by the abundance in beryl of different generations of primary microscopic inclusions of chlorides and fluorides of the alkali metals and of fluorite and carbonic acid (e.g. Cameron, 1953). This fact together with the characteristics of beryllium chemistry, suggests that the process of formation of beryl in pegmatites consists of the dissociation of the mobile beryllium compounds and the fixation of beryllium in solid phase in the form of the difficulty soluble metasilicate of beryllium and aluminium.

Proof that the generally outlined theory is acceptable is demonstrated statistically by Beus (1962) who shows that the beryllium content of the mafic minerals is higher in alkaline rocks compared to granitic rocks, yet beryllium minerals are usually associated with granitic rocks. Beus sites the following average beryllium contents:

TABLE 31

	<u>Granite (ppm)</u>		<u>Alkaline Rocks (ppm)</u>
Muscovite	50	Aegirine	25
Biotite	16	Arvedsonite	30
Hornblende	16	Eudialyte	100

The major minerals of alkaline magmas clearly have a higher concentration of beryllium. The distinctive geochemical features of a crystallizing alkalic magma clearly favour the inclusion of beryllium into silicate lattices. These features are:

- (1) High content of Ti, Zr, rare earths, and Nb.
- (2) Prolonged participation of cations with high valence in the process of mineral formation.
- (3) The alkaline character of the medium which makes the presence of the $(\text{BeO}_4)^{6-}$ complex possible.

It may be noted that X-ray fluorescence scans on pegmatites and aplites on Birch Portage rocks has revealed no high valence cations and it is concluded that the absence of these has largely influenced the retention of beryllium in the melt from which these rocks crystallized.

Date	Particulars	Debit	Credit
1891	To Balance	100	
1892	By Balance		100
1893	By Balance		100

The following is a list of the names of the persons who have been elected to the office of Mayor of the City of New York, from 1891 to 1893, inclusive. The names are given in alphabetical order, and the year of election is given in parentheses.

1891 (1891) - John A. B. (1891)

1892 (1892) - John A. B. (1892)

1893 (1893) - John A. B. (1893)

The following is a list of the names of the persons who have been elected to the office of Mayor of the City of New York, from 1891 to 1893, inclusive. The names are given in alphabetical order, and the year of election is given in parentheses.

1891 (1891) - John A. B. (1891)

1892 (1892) - John A. B. (1892)

1893 (1893) - John A. B. (1893)

APPENDIX III

PROPERTIES OF BERYL

Formula:	$\text{Be}_3\text{Al}_2\text{Si}_6\text{O}_{18}$
Hexagonal:	Uniaxial Negative
	N_p 1.567-1.599
	N_m 1.567-1.608
	$N_m - N_p = 0.004-0.008$
Cleavage:	(001) Poor
Hardness:	7.5-8.0
Streak:	White
Lustre:	Vitreous
Colour:	Colourless, Translucent colour, White, Yellow, or Green.
S.G. _{measured} :	2.65-2.92
S.G. _{calculated} :	0.05-0.10 less than measured
Point Group:	$\frac{6}{m} \frac{2}{m} \frac{2}{m}$
Space Group:	$C \frac{6}{m} \frac{2}{c} \frac{2}{c}$
a_o	$9.20 \pm 0.20 \text{\AA}$
c_o	$9.19 \text{\AA} \pm 0.10 \text{\AA}$
Cell Content:	$Z = 2$

APPENDIX IV

BERYLLIUM MINERALSSilicatesTectosilicates

Epididymite	$\text{Na}(\text{BeSi}_2\text{O}_7)(\text{OH})$
Eudidymite	
Chkalovite	$\text{Na}_2(\text{BeSi}_2\text{O}_6)$
Helvite	$\text{Mn}_8(\text{BeSiO}_4)_6\text{S}_2$
Danalite	$\text{Fe}_8(\text{BeSiO}_4)_6\text{S}_2$
Genthelvite	$\text{Zn}_8(\text{BeSiO}_4)_6\text{S}_2$
Bavenite	$\text{Ca}_4[(\text{Be}, \text{Al})_4\text{Si}_9(\text{O}, \text{OH})_{26}](\text{OH})$

Probably tectosilicates with undetermined structure

Leucophanite	$\text{Na}_{0.6-1}\text{Ca}(\text{BeSi}_2\text{O}_6)$
Meliphanite	$\text{Fe}_{0.6-1}\text{Ca}(\text{BeSi}_2\text{O}_6)$
Aminoffite	$\text{Ca}_2(\text{BeSi}_2\text{O}_6)(\text{OH})_2$
Trimerite	$(\text{Mn}_2, \text{Ca})(\text{BeSiO}_4)_3$
Gadolinite	$(\text{Y}, \text{Ca})_2\text{Fe}(\text{BeSiO}_4)_2(\text{O}, \text{OH})_2$

Metasilicates and Dimetasilicates with Ring Structure

Beryl	$\text{Al}_2\text{Be}_3(\text{Si}_6\text{O}_{18})$
Milarite	$\text{KCa}_2(\text{Be}_2, \text{Al})(\text{Si}_{12}\text{O}_{30})$

Orthosilicates with Nesosilicate Structure

Phenakite	Without additional anions + $\text{Be}_2(\text{SiO}_4)$
Euclase	With OH^- $\text{AlBe}(\text{SiO}_4)(\text{OH})$

Table 1

Summary of the data

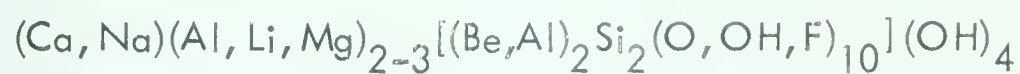
Summary of the data		Source
1. General information		
1.1. Name of the project	1.1.1. Project name	1.1.1.1. Project name
1.2. Date of the project	1.2.1. Date of the project	1.2.1.1. Date of the project
1.3. Location of the project	1.3.1. Location of the project	1.3.1.1. Location of the project
1.4. Duration of the project	1.4.1. Duration of the project	1.4.1.1. Duration of the project
1.5. Budget of the project	1.5.1. Budget of the project	1.5.1.1. Budget of the project
2. Objectives of the project		
2.1. Main objective	2.1.1. Main objective	2.1.1.1. Main objective
2.2. Secondary objectives	2.2.1. Secondary objectives	2.2.1.1. Secondary objectives
2.3. Specific objectives	2.3.1. Specific objectives	2.3.1.1. Specific objectives
2.4. Expected results	2.4.1. Expected results	2.4.1.1. Expected results
3. Methodology of the project		
3.1. Research methodology	3.1.1. Research methodology	3.1.1.1. Research methodology
3.2. Data collection methodology	3.2.1. Data collection methodology	3.2.1.1. Data collection methodology
3.3. Data analysis methodology	3.3.1. Data analysis methodology	3.3.1.1. Data analysis methodology
3.4. Quality control methodology	3.4.1. Quality control methodology	3.4.1.1. Quality control methodology
4. Results of the project		
4.1. Main results	4.1.1. Main results	4.1.1.1. Main results
4.2. Secondary results	4.2.1. Secondary results	4.2.1.1. Secondary results
4.3. Specific results	4.3.1. Specific results	4.3.1.1. Specific results
4.4. Expected results	4.4.1. Expected results	4.4.1.1. Expected results
5. Conclusions of the project		
5.1. Main conclusions	5.1.1. Main conclusions	5.1.1.1. Main conclusions
5.2. Secondary conclusions	5.2.1. Secondary conclusions	5.2.1.1. Secondary conclusions
5.3. Specific conclusions	5.3.1. Specific conclusions	5.3.1.1. Specific conclusions
5.4. Expected conclusions	5.4.1. Expected conclusions	5.4.1.1. Expected conclusions

Diorthosilicates with Nesosilicate Structure

Barylite	Without additional anions $\text{BaBe}_2(\text{Si}_2\text{O}_7)$
Bertrandite	With OH^- $\text{Be}_4(\text{Si}_2\text{O}_7)(\text{OH})_2$
Gelbertrandite	$\text{Be}_4(\text{Si}_2\text{O}_7)(\text{OH})_2 \cdot n\text{H}_2\text{O}$
Spherobertrandite	$\text{Be}_5(\text{Si}_2\text{O}_7)(\text{OH})_4$
Beryllite	$\text{Be}_5(\text{Si}_2\text{O}_7)(\text{OH})_4 \cdot 2\text{H}_2\text{O}$

Beryllium Aluminosilicates with Phyllosilicate Structure

Beryllium margarite-bityite-bityite-bowleyite group



Oxides and Multiple Oxides

Bromellite	BeO
Chrysoberyl	Al_2BeO_4
Taaffeite	$\text{Al}_4\text{MgBeO}_8$

Borates

Hambergite	$\text{Be}_2(\text{BO}_3)(\text{OH})$
Rhodizite	$(\text{K}, \text{Na})_{1-2}(\text{Li}, \text{Al})_{1.5-4}\text{Al}_4[(\text{Be}, \text{B})_3\text{B}_4\text{O}_{12}]_2(\text{O}, \text{OH})_4$

Antimonates

Swedenborgite	$\text{NaBe}_4\text{SbO}_7$
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Phosphates

Beryllonite	$\text{NaBe}(\text{PO}_4)$
Hurlbutite	$\text{CaBe}_2(\text{PO}_4)$
Herderite	$\text{CaBe}(\text{PO}_4)(\text{OH}, \text{F})$
Vayrynenite	$\text{MnBe}(\text{PO}_4)(\text{OH})$
Moraesite	$\text{Be}_2(\text{PO}_4)(\text{OH}) \cdot 4\text{H}_2\text{O}$
Kolveckite	$\text{Ca}_{0.1}(\text{Al}, \text{Be})[(\text{P}, \text{Si})\text{O}_4] \cdot 2\text{H}_2\text{O}$

Carbonates

Beryllium tengerite	$(\text{Y}, \text{Ce})\text{Be}(\text{CO}_3)(\text{OH})_3 \cdot \text{H}_2\text{O}$
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APPENDIX V

Specimens in Order of Description in Chapter 3

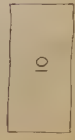
560	Porphyroblastic Biotite Gneiss (Unit 1)
587	Porphyroblastic Hornblende Biotite Gneiss (Unit 2)
550	Porphyroblastic Hornblende Plagioclase Gneiss (Unit 3)
527	Hornblende K-feldspar Gneiss (Unit 4)
599	K-feldspar Gneiss (Unit 4a)
598	Plagioclase Amphibole Gneiss (Unit 5)
530	Greenish-black Basic Dyke (Unit 6)
513	Black Basic Dyke (Unit 6)
515	Porphyritic Felsite Dyke (Unit 7)
503	Feldspar Porphyry Dyke (Unit 8)
501	Cream Aplite Dyke (Unit 9)
519	Cream Aplite Dyke (Unit 9)
510	Red Aplite Dyke (Unit 9)
533	Red Aplite Dyke (Unit 9)
588	Barren Pegmatite Dyke with Xenoliths (Unit 10)
556	Barren Pegmatite Dyke with parallel crystal growth (Unit 10)
526	Rhythmically Banded Barren Pegmatite Dyke (Unit 10)
531	Barren Pegmatite Dyke (Unit 10)
539	Barren Pegmatite Dyke (Unit 10)
556	Barren Pegmatite Dyke with parallel crystal growth (Unit 10)
601A	Rhythmically Banded Beryl Pegmatite Dyke (Unit 10)
549	Beryl Pegmatite Dyke with parallel crystal growth (Unit 10)
512	Sodic Beryl Pegmatite Dyke (Unit 11)
521	Beryl Pegmatite Dyke with primary Spessartite (Unit 11)

CONTENTS OF VOLUME I

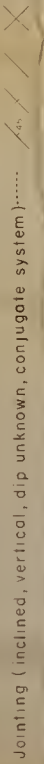
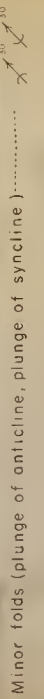
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Chapter XXV. The History of the	495
Chapter XXVI. The History of the	515
Chapter XXVII. The History of the	535
Chapter XXVIII. The History of the	555
Chapter XXIX. The History of the	575
Chapter XXX. The History of the	595

- 522 Beryl Pegmatite Dyke with yellow Beryl (Unit 11)
- 601 Rhythmically Banded Beryl Pegmatite Dyke (Unit 11)
- 608 Central phase Perthite with twinned Albite (Unit 10)
- 608 As above under lower power (Unit 10)
- 601A Central Phase Perthite with untwinned Albite (Unit 11)

LEGEND

	Beryl pegmatite
	Barren pegmatite
	Aplite
	Feldspar porphyry
	Porphyritic felsite
	Basic dyke
	Plagioclase amphibole gneiss
	Hornblende K-feldspar gneiss; 4a lenses of fine grained K-feldspar gneiss in 4, 4b lenses of amphibolite in 4
	Porphyroblastic hornblende plagioclase gneiss
	Porphyroblastic hornblende biotite gneiss; 2a porphyroblastic hornblende biotite granulite; 2b porphyroblastic hornblende biotite migmatite
	Porphyroblastic biotite gneiss; 1a porphyroblastic biotite granulite; 1b porphyroblastic biotite migmatite

SYMBOLS

	Outcrop area
	Geological boundary (defined, approximate, assumed)
	Dyke (inclined, vertical, dip unknown)
	Pegmatite body (plug-like, sheet-like)
	Foliation (inclined, vertical)
	Mineral lineation
	Drag folds (z-drag, s-drag, w-drag)
	Jointing (inclined, vertical, dip unknown, conjugate system)
	Minor folds (plunge of anticline, plunge of syncline)
	Specimen location
	Winter road or trail

Geology by D. Radcliffe, 1963
To accompany Report No.

Base map prepared from enlargement of R.C.A.F. one quarter mile to one inch scale vertical aerial photograph A 9788-43.

102°40' 20"

54°54' 08"

54°52' 15"

102°40' 20"

THE BIRCH PORT

102° 37' 8" 54° 54' 08"

ONIKUP LAKE

WETTON LAKE

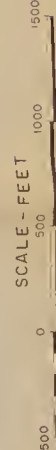
WELLS LAKE

STURGEON WEIR RIVER

102° 37' 38" 54° 52' 15"

PORTAGE BERYL PEGMATITE DEPOSIT

SASKATCHEWAN



B29823